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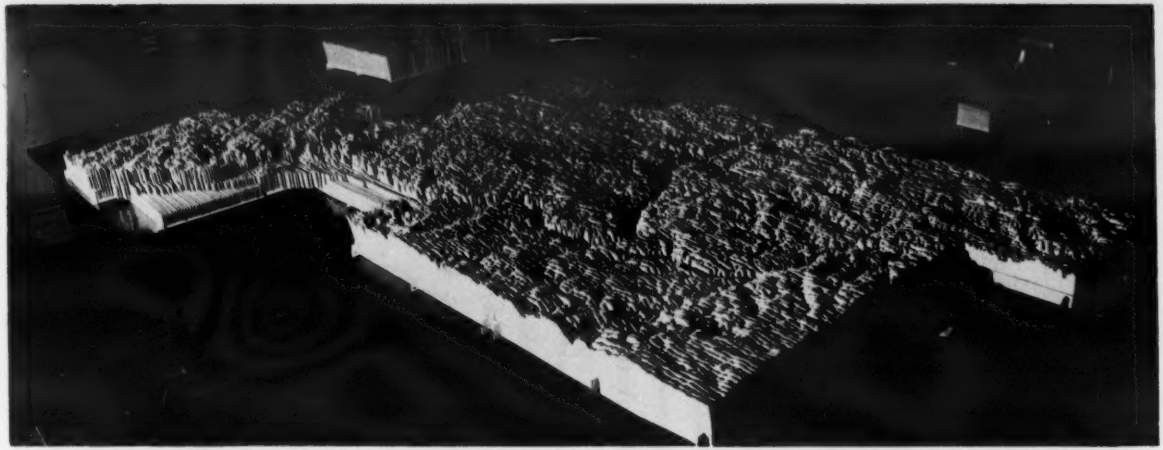

FIGURE 1. SOUTHWESTERN PART OF DRIFTLESS AREA AND OF WAUSAU-MARATHON AREA OF NORTHERN WISCONSIN
Looking northeast. Location given in insert in Figure 1.

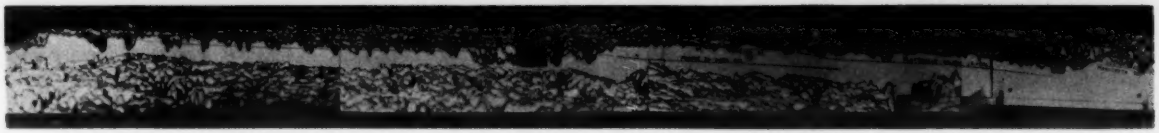

FIGURE 2. PART OF PROFILES REMOVED
To show relation of surface to structure along a N – S section. Looking eastward. Flattening of structure near crest of Platteville-Galena cuesta in right center of section is local and not typical of crests of cuestas. *See* Plate 2, figure 2.

PROJECTED PROFILES OF THE DRIFTLESS AREA

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GEOMORPHIC HISTORY OF THE KICKAPOO REGION, WISCONSIN

BY ROBERT E. BATES

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ABSTRACT

The Kickapoo region, located in the Driftless Area of southwestern Wisconsin, consists of a submaturely dissected plateau developed on gently dipping sedimentary rocks.

The upland of the region has been variously interpreted as: (1) a series of single-cycle cuestas, (2) remnants of one peneplane on the lower parts of the upland, and (3) two peneplanes, one, that mentioned above, and another on the higher parts of the upland. From field observation and laboratory study of projected profiles the higher parts of the upland are interpreted as remnants of one peneplane—the Dodgeville of Trowbridge—thus eliminating the first two of the above interpretations. The lower parts of the upland, while possibly remnants of a partial peneplane, may equally well be interpreted as structural surfaces. These two conclusions constitute a fourth interpretation.

The Kickapoo River is a strikingly misfit stream. In explaining the misfit relation, consideration is given the mechanics of meander development both in floodplain and

incised streams. Both theoretical considerations and data from topographic maps indicate that for streams of a given size, those that are incised have a larger maximum meander size than those flowing on floodplains. Consequently, a misfit stream may result from the sudden change from incised to floodplain characteristics. This, as well as climatic change and underflow through the alluvium, is a possible explanation for the misfit character of the Kickapoo.

The extensive alluvial deposits of the Kickapoo Valley, almost entirely the result of slope wash, probably result from aggradation of the Wisconsin Valley to which the Kickapoo is tributary.

INTRODUCTION AND ACKNOWLEDGMENTS

The Kickapoo region of southwestern Wisconsin centers about, and for the most part is included in, the drainage basin of the Kickapoo River. The Kickapoo is a south-flowing tributary of the Wisconsin River and the largest stream that has its entire drainage basin within the Driftless Area (Fig. 1).

In this paper the geomorphic history of the Kickapoo region is considered in two parts. The first part concerns the development of the upland surface, below which the streams have intrenched themselves. Most of the Driftless Area will be included in this discussion. The second part considers the development of the present valleys of the Kickapoo River system. Development of the upland includes a study of the peneplane problem, concerning which widely divergent views have been expressed, and also the relation of the upland of this region to the supposed pre-Cambrian peneplane of northern Wisconsin. Field work on these problems has been supplemented by laboratory study of a series of projected profiles. Consideration of the development of the valleys includes a study of the misfit character of the streams, rock terraces present in parts of the valleys, and deposits and forms resulting from the aggradation of the valleys. The Kickapoo River has been affected by glaciation only insofar as conditions in the Wisconsin River, an outlet for glacial meltwaters, reacted upon it. Consequently, most of the complications found in streams in glaciated regions are lacking in the Kickapoo Valley.

The study, undertaken at the suggestion of Professor Douglas Johnson, has been carried out under his supervision. The writer wishes to express sincere appreciation of his friendly advice and helpful criticism.

Acknowledgment is due M. King Hubbert of Columbia University for his valuable assistance in formulating the analysis of meander growth.

The writer is indebted to other workers in the field, A. C. Trowbridge of the State University of Iowa, F. T. Thwaites of the University of Wisconsin, Lawrence Martin, formerly of the University of Wisconsin, and now Chief of the Division of Maps, Library of Congress, and Paul MacClintock of Princeton University, for their cordial assistance both in

discussing some phases of the problem and in extending the use of facilities and equipment which assisted materially in the work.

The construction of projected profiles of the region was made possible

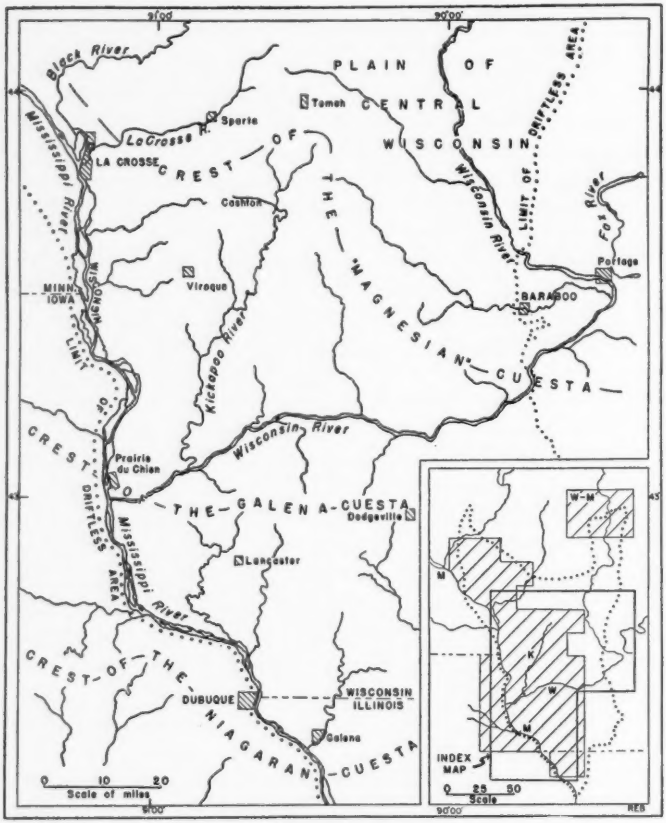

FIGURE 1.—*Index map*

Driftless Area outlined in insert by dots. Areas for which projected profiles have been constructed, including the Wausau-Marathon area (W-M), are shown by diagonal lines. Mississippi River (M), Wisconsin River (W), Kickapoo River (K).

in part by grants from the Marsh Fund of the National Academy of Sciences, the Federal Emergency Relief Administration, and the National Youth Administration.

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versity, Gilbert O. Raasch and Carl A. Bays of the University of Wisconsin, and George I. Whitlatch of the Tennessee Geological Survey for their helpful suggestions while in the field with the writer, and to Elliot H. Powers of the State University of Iowa for structural data on parts of the region.

Finally, the writer is indebted to Henry S. Sharp and Ralph L. Miller of the faculty, and to members of the seminar in Geomorphology at Columbia University for critical suggestions made both in the field and in the laboratory.

GEOLOGIC AND TOPOGRAPHIC RELATIONS

GEOLOGIC RELATIONS

The geomorphic development of the Kickapoo region has been influenced to a remarkable extent by the stratigraphy and structure of the underlying formations. These factors must, therefore, be considered in any interpretation of the surface features. The rocks of the region consist of Lower Paleozoic formations lying on the southwestern side of the Wisconsin dome. The sequence is as follows:

Stratigraphic Units	Thickness (Feet)	Geomorphic Expression
Silurian		
Niagaran dolomite	150	Resistant cuesta-maker
Ordovician		
Maquoketa shale	200	Nonresistant lowland-maker
Galena dolomite Decorah shale Platteville limestone	200–300	Resistant cuesta-maker
St. Peter sandstone		Nonresistant lowland-maker
Unconformity	330	
Prairie du Chien formation Oneota dolomite 75–150		Resistant cuesta-maker
Cambrian		
St. Croixan sandstone	800–900	Nonresistant lowland-maker
Pre-Cambrian		
Crystallines		Resistant upland-maker

The less resistant St. Croixan sandstone underlies the extensive low plain of central Wisconsin, just northeast of the Kickapoo basin. Only the Oneota dolomite member of the Prairie du Chien group is present in the region studied. This formation is responsible for the prominent northeastward-facing Prairie du Chien or "Magnesian cuesta" (Fig. 1),

overlooking the plain of central Wisconsin; the Kickapoo basin is located on its backslope. Because of the unconformity above the Oneota, its thickness in this area ranges from 75 to 150 feet, but the combined thickness of the Oneota and the St. Peter is more nearly constant at approximately 330 feet.

The St. Peter sandstone is pure white, occasionally iron-stained, and saccharoidal. The topmost 2 or 3 feet is very strongly cemented with iron oxide and is distinctively resistant. Its line of outcrop is consistently characterized by a pronounced shelf.

The resistant layer at the top of the St. Peter, together with the overlying Platteville, Decorah, and Galena has a combined thickness of 200 to 300 feet. As a unit they hold up a prominent cuesta of the region (Fig. 1). South of the Wisconsin River the cuesta is known as Military Ridge.

The next succeeding Maquoketa formation is approximately 200 feet thick. Predominantly shale, and underlying an inter-cuesta lowland, it forms a nonresistant unit. The highest formation exposed in the Driftless Area is the Niagaran dolomite [1] which varies in thickness up to 150 feet (Sutton, 1935, p. 268). As is true of the other limestones and dolomites in the section, this resistant formation forms a prominent cuesta (Fig. 1).

The formations of the southern part of the Driftless Area, including the Kickapoo region, dip slightly west of south, away from the pre-Cambrian dome of northern Wisconsin. In the northeastern part of the Driftless Area, however, the dip is more nearly west. In most places it is less than 15 feet per mile and becomes as much as 30 feet per mile only in the extreme southern part. Minor structural irregularities are not reflected in the topography. Because of the slight dip of the formations the belts of outcrops, which are roughly concentric around the dome, are extremely ragged and irregular.

TOPOGRAPHIC RELATIONS

Except in the Baraboo region, the Driftless Area at the present time is a submaturely dissected plateau. Below its surface lie the incised valleys of the present streams and the broad lowland of central Wisconsin, both of which are developed on the weak St. Croixan sandstones. The surface of the plateau slopes southward from an elevation of more than 1400 feet just south of Sparta and Tomah to 800 or 900 feet in northwestern Illinois. As a whole, this upland is remarkably even and contrasts sharply with the valleys incised as much as 500 feet below it.

[1] While commonly called the Niagaran dolomite, this is not stratigraphically correct since it is in part Alexandrian, older than Niagaran, and possibly in part Cayugan, younger than Niagaran.

Detailed examination of the upland, however, reveals numerous irregularities in the generally even surface. Broad flat-topped ridges or cuestas mark the line of outcrop of the resistant Prairie du Chien, Platteville-Galena, and Niagaran formations. Between these roughly parallel ridges, the upland is somewhat lower along the outcrop of the less resistant formations. Thus the upland consists of a series of low cuestas and intervening lowlands several hundred feet above the present streams incised in narrow valleys. The valley of the Wisconsin River in part of its course across the Driftless Area is cut below the intercuesta lowland developed on the St. Peter sandstone. The development of this valley has largely destroyed the lower part of the upland surface in the surrounding regions. This lower part of the upland developed on the St. Peter is preserved in other areas as will be discussed later.

DEVELOPMENT OF THE UPLAND

PENEPLANATION

Before discussing the Kickapoo River it is well to consider the development of the upland below which the Kickapoo is intrenched. The principal question concerns peneplanation—namely, whether the present upland consists wholly or in part of peneplane remnants, and, if so, how many peneplanes are represented. Because peneplanation is a regional problem, such a study cannot be satisfactorily limited to one drainage basin. The entire southern and western parts of the Driftless Area are, therefore, included in the study. This region is particularly favorable for such a study for two reasons: (1) the topography, unlike that of the surrounding region, has not been buried by deposits of continental glaciation; and (2) topographic maps of the region have been recently completed, greatly facilitating the work and serving as a basis for the construction of projected profiles.

PROJECTED PROFILES: THEIR USE AND CONSTRUCTION

In regions of horizontal or gently dipping strata of varying resistance, stripped structural surfaces in some cases can be distinguished from peneplanes only with the greatest difficulty. The distinguishing features become less pronounced as the dip of the strata approaches more closely the gradient and the direction of flow of the major graded streams of the area. This difficulty is encountered to a great degree in the Driftless Area. The extremely gentle dip, in most places less than 15 feet per mile, is in approximately the same direction and only slightly greater than the general slope of the upland. Both are slightly more than the gradient of the present graded streams, many of which flow in the direction of dip. Because of the close accordance of the slope of the upland

surface and the dip of the underlying strata, the features of this area have been interpreted by some as a series of cuestas, by others as peneplane remnants. Application of these two interpretations to the same area must be expected in areas exhibiting close parallelism of topography and structure.

To aid in overcoming the difficulty presented by such close parallelism, multiple projected profiles have been constructed of a large part of the Driftless Area, as shown in Figure 1. The advantage of the method of multiple projected profiles, developed at Columbia University for the study of peneplane problems, is that it not only makes possible the studying of the topographic features of an area on a much reduced horizontal scale and on an exaggerated vertical scale which brings out differences in slope too small to be seen in the field, but it also enables the investigator to view the relation of topography to structure from every possible point of view. This method has the additional advantage of being entirely objective, unless the choice of the scale and the spacing of the profiles is considered otherwise.

The method of constructing the profiles consists of mounting the topographic maps of the area to be profiled on cloth in strips of convenient width, for example, the width of one 15-minute quadrangle. The long direction of the strips is made parallel to that in which the profiles are to be drawn. The direction usually chosen is at right angles to the trend of the structural and topographic features to be studied. These strips of mounted maps are then ruled lengthwise at intervals of two inches, one inch, or half an inch depending on the detail desired. A profile of each of these narrow strips is then constructed on cross-section paper by plotting the highest contour within the strip at intervals of one-tenth of an inch along the strip, and then joining the successive points with a line. The profile constructed in this way does not represent the profile along any single line, but rather that of the highest contours throughout the narrow strip. The profile is then traced on cardboard, cut out with a jig-saw, and finally, is lined along both sides of the upper edge with a heavy black line. In order to study the profiles they are placed upright in their proper positions in slots cut in cardboard or wooden strips.

In the Driftless Area the trend of the structure and topography is slightly north of west, and the profiles were accordingly constructed in a north-south direction. Here, because of the apparent importance of smaller details in the topography, an interval of half an inch (half a mile) was chosen for the spacing of the profiles. Each profile thus represents the highest points within a belt half a mile wide north and

south across the area profiled. A vertical exaggeration of 26.4 times was selected as sufficient to bring out minor differences in slope, at the same time to retain the general uniformity of surface which is so striking as

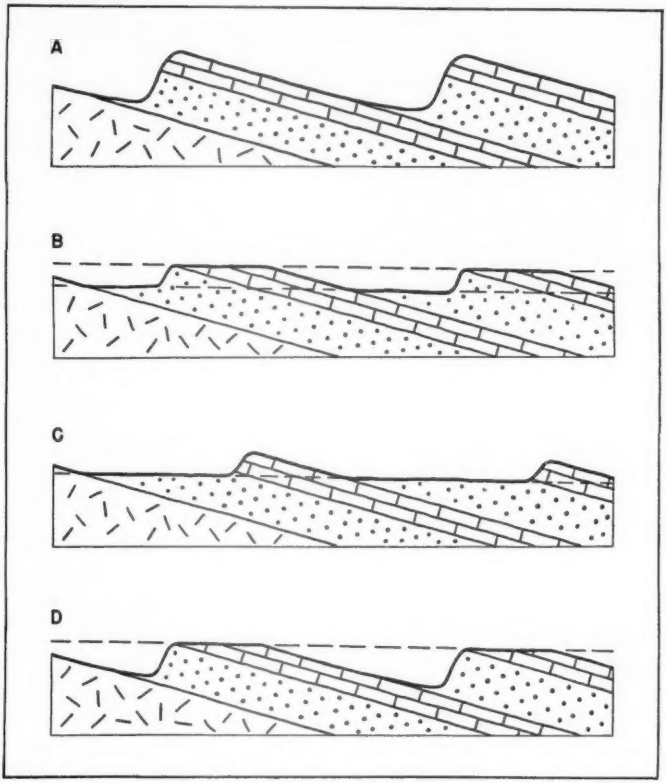


FIGURE 2.—*Alternative upland histories*

A. Series of cuestas without peneplanation (Martin).
B. Two peneplanes (Trowbridge).
C. One peneplane developed on lower parts of upland (Thwaites).
D. One peneplane developed on higher parts of upland (Bates).

seen in the field. To bring out more clearly the relations of topography and structure, lines were drawn along each cardboard at the contact of the pre-Cambrian and the St. Croixan, the St. Croixan and the Prairie du Chien, the St. Peter and Platteville, and the Maquoketa and Niagaran. The structural data have been assembled from various publications

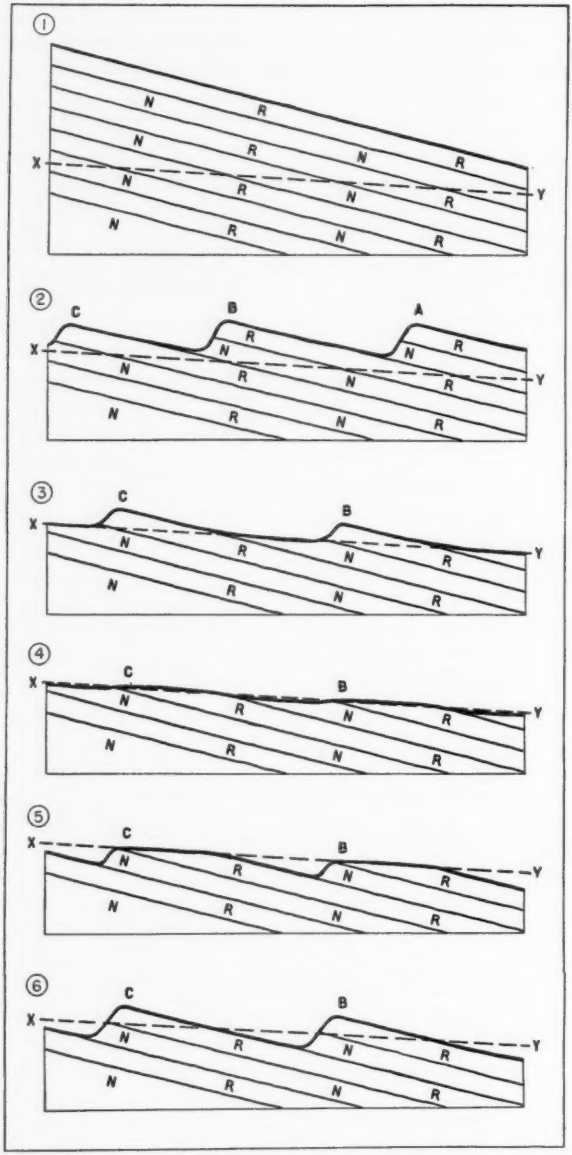


FIGURE 3.—*Deduced results of erosion on tilted beds of varying resistance*

R and N represent resistant and nonresistant strata, respectively. Cuestas A, B, and C are developed as in 2; reduced to a peneplane at the level XY as in 3 and 4; 3 and 4 are then rejuvenated in 6 and 5, respectively.

(Trowbridge, 1934, 1935; Shaw and Trowbridge, 1916; Grant and Burchard, 1907) and from Elliot Powers (personal communication).

PREVIOUS INTERPRETATIONS

General review.—The erosional history of the Driftless Area has been variously interpreted, ranging from the development of all the features in one cycle (Martin, 1932, p. 69-77) to the other extreme in which five erosion levels have been identified above the present streams (Hershey, 1896, p. 72-100). Only three interpretations treat any considerable portion of the Driftless Area, and these three alone are given detailed consideration in this paper. The first was that proposed by Martin (1932, p. 69-77); according to him all features of the Driftless Area have been developed by differential erosion during the present cycle (Fig. 2A). Trowbridge (1921, p. 126-127) interprets the upland as remnants of two peneplanes (Fig. 2B). Thwaites (1935, p. 120) has recently advanced a third "compromise" interpretation favoring the existence of one peneplane (Fig. 2C).

Martin's hypothesis of one-cycle cuestas.—Martin is of the opinion that by subjecting a succession of gently dipping, alternately resistant and nonresistant strata to differential erosion it is possible to produce in the present cycle all the features seen today in the Driftless Area. He (Martin, 1932, p. 77) therefore

> "inclines toward the simpler explanation, that of cuestas, unless some conclusive evidence of peneplanation can be found."

In support of this interpretation it may be stated that the higher parts of the upland are with few exceptions along the northernmost limits of outcrop of the resistant limestones and dolomites, and the lower parts of the upland are between the ridges composed of the resistant formations.

A theoretical analysis of the topographic forms developed in two cycles of erosion on a series of gently dipping, alternately resistant and nonresistant strata is shown in Figure 3, in which for convenience the dip of the beds is represented as much steeper than is the case in the Driftless Area. On the beds shown in Figure 3 (1), a series of cuestas have been developed by differential erosion (Fig. 3 (2)). Practically the entire thickness of the cuesta-maker is preserved to the crests of the cuestas, and most of the backslope is a stripped structural surface. A thin veneer of the overlying weak rock may, however, extend a short distance up the backslope. The crests of these cuestas may or may not rise to accordant elevations. This is determined by a number of factors of which the thickness and relative resistance of the cuesta-makers and the intervening weak rock are among the more important. In the earlier stage shown in Figure 3 (3), the surface is sufficiently well developed

to bevel the weaker rock, but the more resistant crests of the cuestas are reduced only slightly. In the later stage (Fig. 3 (4)), both the weak and the resistant beds are beveled by the surface. If the entire region is reduced to near baselevel as in Figure 3 (4), the resulting peneplane will not be a geometrically plane surface. On the contrary, the topography will reflect in a subdued way the relative resistance of the underlying strata. Figure 3 (6) shows that rejuvenation of such a peneplane as is shown in Figure 3 (3) will result in the removal of weak rock in much the same way as in the development of the cuestas in Figure 3 (2). Cuestas developed following complete peneplanation, however, differ from those developed in a first cycle in that their crests are distinctly beveled, whereas in the first cycle almost the full thickness of the cuesta-maker is preserved to the crests of the cuestas. Farther down the backslope stripped structural surfaces are found in both cases. Assuming the same dip and thickness of formations, it also follows that the scarps of the first cycle cuestas will be considerably higher than those of the second cycle. It should be noted that the cuestas developed by rejuvenation of a partial peneplane such as is shown in Figure 3 (3) would show no beveling along the crests (Fig. 3 (6)). Evidence of peneplanation shown in Figure 3 (5) would be destroyed if the beveled crests were removed by retreat of the scarps.

The conclusions to be drawn from this analysis are that distinct beveling of cuesta crests indicates definitely that such crests are peneplane remnants, but that the absence of beveled crests does not eliminate the possibility of peneplanation of weaker beds between the cuestas or of peneplanation of the cuestas followed by subsequent scarp retreat destroying all evidence of the beveling.

With these deductions in mind we are in a position to examine the conditions existing in the Driftless Area. By field study it is impossible to detect any change in slope from a beveled crest to a stripped structural surface on the backslope of the cuestas. In view of the near parallelism of surface and structure this is not surprising.

The vertical exaggeration of 26.4 times used in constructing the projected profiles is sufficient to show distinct beveling of the cuesta crests and a definite break in slope between the beveled tops of the cuestas and the stripped backslopes (Pl. 1, fig. 2; Pl. 2, figs. 1, 2). The extent of these beveled areas is shown in Figure 4. As would be expected, the beveled areas are narower where the dip is greater—namely, along the crest of the Niagaran escarpment.

Thwaites (personal communication) has raised the question as to whether the beveling of the dolomite cuestas may not be due to the fact that the crests have been exposed to weathering and erosion longer than the areas farther down the backslope which were progressively exposed

by continued erosion. If this were the complete story one would expect a gradual transition from the long-exposed areas on the crests to the areas exposed for a shorter time. However, the profiles reveal not a gradual

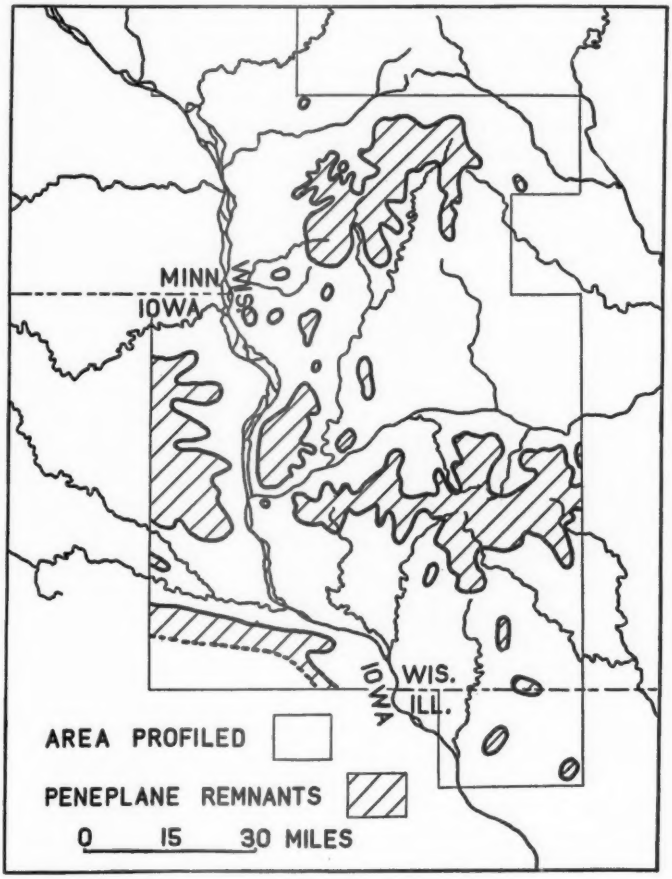


FIGURE 4.—*Map of peneplane remnants*

transition but a definite break between the beveled crests and the stripped backslopes.

Beveling of the structure is also shown along the Kickapoo-Mississippi divide which forms an almost continuous bridge of higher land connecting the crests of the Prairie du Chien and Platteville-Galena cuestas. The

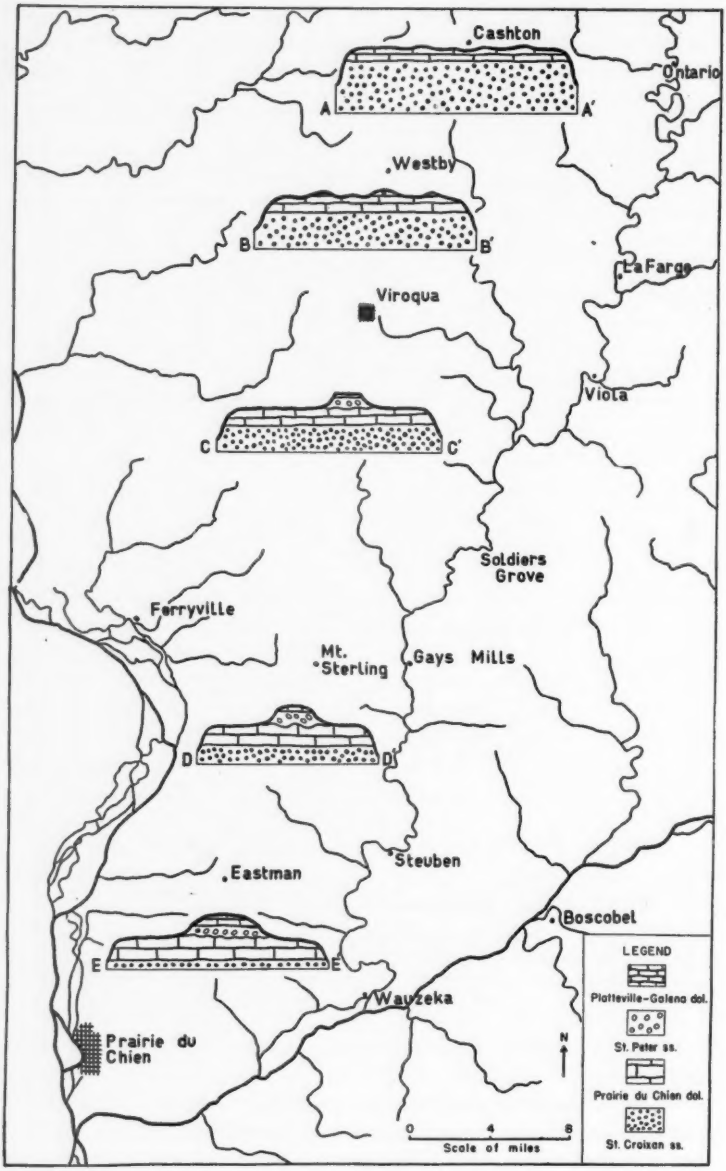


FIGURE 5.—*Sections across Kickapoo-Mississippi divide*

significance of this connecting bridge was noted by Trowbridge (1921, p. 71-72). Throughout most of the length of the divide there are broad flat summit areas which owe their preservation to the favorable location of resistant strata. All of the areas beveling the structure on the cuesta crests and the principal divides can be projected to similar adjoining areas to form a single relatively even surface. It is concluded, therefore, that remnants of one peneplane are present in the area. The peneplane is essentially the one to which Trowbridge (1921, p. 64) has applied the name Dodgeville, from the town of Dodgeville, Wisconsin, where it is well developed. It is to be expected that the parts of the peneplane surface near the divides will be somewhat higher than those along the streams, but the fact that the resistant cuesta crests are so effectively beveled strongly suggests that there was little relief on the peneplane surface.

Trowbridge's hypothesis of two peneplanes.—In addition to the Dodgeville peneplane, Trowbridge has called attention to what he believes is another peneplane about 200 feet lower. This, to which he has applied the name Lancaster [2] from Lancaster, Wisconsin, is considered to be much more extensively represented in the present topography than is the Dodgeville. He includes in it all the upland except the cuesta crests and a few of the principal divides.

Field study reveals that the sharpness of the contact of these two surfaces varies considerably. In some places the contact is almost imperceptible and in others it is marked by a very sharp break in the topography. A north-south traverse along the Kickapoo-Mississippi divide reveals interesting variations in the nature of this contact (Fig. 5). Near the crest of the Prairie du Chien cuesta, in the vicinity of Cashton, the broad flat summit areas (remnants of the Dodgeville peneplane) that bevel the structure break away abruptly into ravines leading to the present stream level with no intermediate bench or widening of the valleys that might be interpreted as a lower peneplane Cross-section AA′). A little farther south the divide is somewhat more irregular, the higher parts, underlain by St. Peter sandstone, do not form a recognizable erosion level. Extensive areas at slightly lower elevations on both sides of the divide are developed on the Prairie du Chien dolomite (Cross-section BB′). Still farther south, near Viroqua, the resistant series consisting of the upper few feet of the St. Peter and the overlying Platteville-Galena sequence favors the preservation of broad flat summit areas. The resistant series underlying these summit areas consists of a wedge which thickens southward. Just south of Viroqua where this wedge is thin, the crest of the

[2] The term "Lancaster peneplane" was first applied to a more limited area by Grant and Burchard (1907).

divide is flat as shown in Cross-section CC′, and the break from the upper flat to the more extensive lower areas on either side of the divide underlain by Prairie du Chien dolomite is very abrupt. Progressing southward, as the wedge of Platteville-Galena becomes thicker, the divide is somewhat more rounded and the topographic break at the St. Peter-Platteville contact is less abrupt (Cross-section DD′). Whatever break there is in slope still occurs at the same horizon—the top of the St. Peter—but as the horizon occupies a lower and lower position on the slope, the break is less conspicuous and almost merges with the general slope (Cross-section EE′). The lower parts of the upland on each side of the divide are still developed on the Prairie du Chien, but the difference in elevation between these lower parts and the upper surface along the crest of the divide becomes progressively greater toward the south.

The more or less conspicuous summit areas, accordant in elevation from the crest of one cuesta to another, occur on successively higher stratigraphic horizons toward the south. At the same time, the lower parts of the upland are lacking near the crest of the cuesta and occur at increasingly greater distances below the upper areas as one goes southward. This suggests that the lower surfaces here developed on the Prairie du Chien are of structural origin. Study of the projected profiles of the area reveals that the surface of the lower areas developed on the Prairie du Chien and the surface developed on the Platteville-Galena coincide closely with the dip of the strata (Pl. 2, figs. 1, 2). Similar conditions are strongly suggested for the surface on the Niagaran dolomite, but the small area of Niagaran outcrop included in the profiles is insufficient basis for a definite statement. All evidence from field study and from study of profiles of the Driftless Area establishes the existence of one peneplane, but no evidence from either of these sources definitely points to the existence of a second peneplane at a lower level.

If the flat summit areas are remnants of the only peneplane in the region, all features of the topography below that general level must have been developed in a period or periods too short for general reduction of the country well toward a new base level. The fairly extensive lower surfaces of the upland, parallel to the structure, must have developed for the most part either while the streams flowed approximately at the level of those surfaces or while they flowed some distance below them.

If the streams flowed at any particular level for sufficient time to permit the development of stripped structural surfaces at that level, they must have been held there by their proximity either to sea level or to the level of a resistant barrier which constituted a local base level. The form resulting from the first of these conditions would be of cyclical significance. That developed by the second condition might be hard

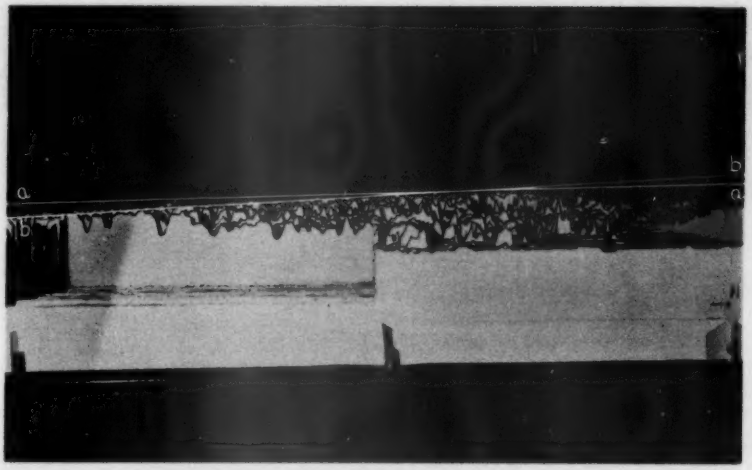


FIGURE 1. CREST OF PRAIRIE DU CHIEN CUESTA
Looking westward. In right foreground is low plain of central Wisconsin. Skyline in right center shows beveled crest of Prairie du Chien cuesta (a, a), representing Dodgeville peneplane, which is also shown along Kickapoo-Mississippi divide in left distance. In left foreground can be seen stripped structural surface developed on backslope of cuesta (b, b).

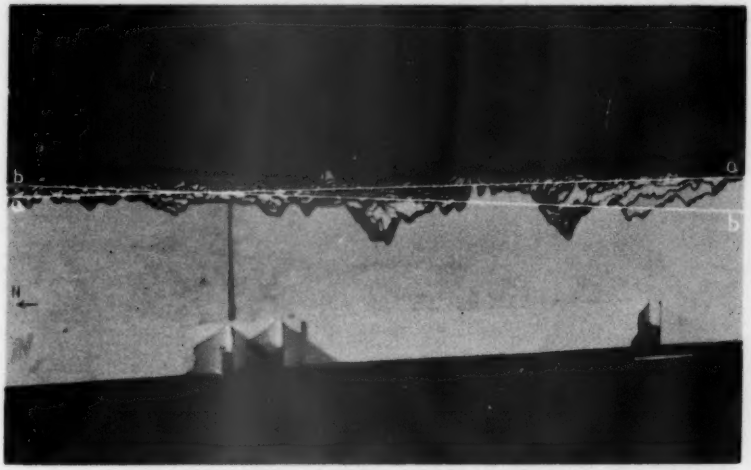


FIGURE 2. BACKSLOPE OF PLATTEVILLE-GALENA CUESTA
The nearly horizontal peneplane remnants bevel crests of Platteville-Galena (left) and Niagaran (right) cuestas (a, a). Between cuestas is southward (toward the right) sloping stripped structural surface on backslope of Platteville-Galena cuesta (b, b).

PROJECTED PROFILES OF BEVELED CUESTA CRESTS

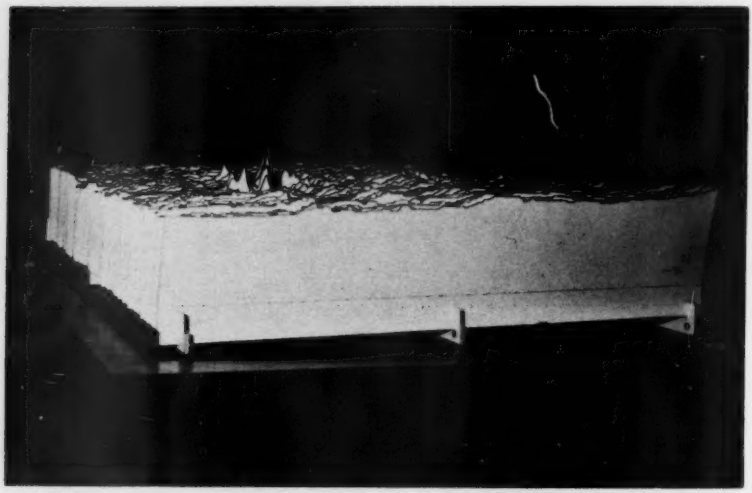


FIGURE 1. LOOKING WEST-NORTHWEST
The nearly horizontal bench is seen in northeastern corner (right) of area and the more regularly sloping surface along western (distant) edge of profiles. Prominent peaks in center are underlain by Huronian quartzite.

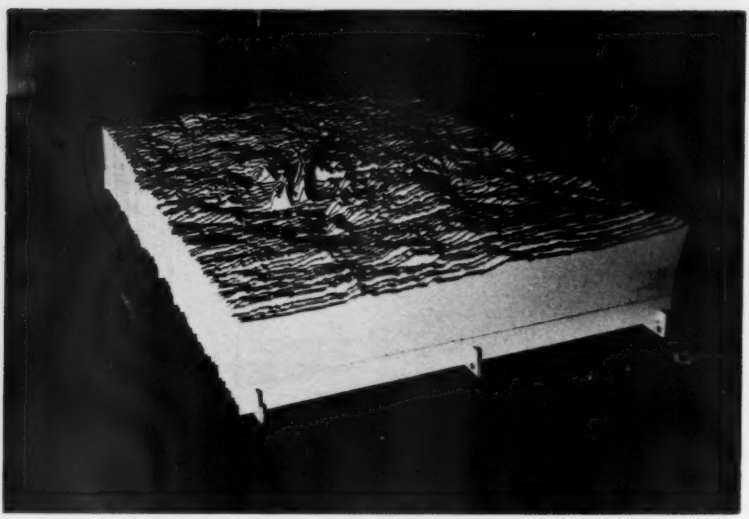


FIGURE 2. A GENERAL VIEW
Highest peak near center is Rib Hill.

PROJECTED PROFILES OF THE WAUSAU-MARATHON AREA

to distinguish from the first after rejuvenation. Although surfaces similar to the lower parts of the upland of the Driftless Area might be the result of a partial erosion cycle, the existence of such a cycle can be verified only by the presence of remnants of the resulting surface that distinctly bevel the structure. As has been pointed out, no such remnants of a second peneplane are found here, and it is therefore concluded that the extensive lower parts of the upland do not necessarily have cyclical significance.

One may question whether such surfaces can be developed some distance above stream level. It seems probable that the features under discussion are related to the prominently developed bench topography of semi-arid plateau regions. The more striking appearance of the bench topography in those regions may be explained by the paucity of vegetation and the occurrence of precipitation in the form of cloudbursts which greatly accelerate erosion by rill action and slope wash. If such surfaces are formed some distance above the principal streams it must be assumed that the resistance of the stratum on which the surface is developed is sufficient to maintain a raised local base level long enough to permit the overlying weak rock to be removed. Fenneman (1936, p. 181) has expressed the belief that

> "the strong stratum may actually raise the local base level a little for a long time, or may raise it much for a short time, but not much for a long time."

He is undoubtedly correct that, other things being equal, the more a local base level is raised by a strong stratum the more rapidly the barrier will be cut away. It does not necessarily follow, however, nor is it implied by Fenneman, that a local base level can never be maintained for sufficient time to permit the removal of the overlying weak rock. The absolute length of time which a raised local base level can be maintained by a resistant stratum is dependent on a number of factors. Among the more important of these are the thickness and relative resistance of the various strata involved, the extent of the vegetative cover, and the amount and character of precipitation. Whether or not the particular conditions that existed in the Driftless Area would have permitted the development of benches is, in the absence of critical data, obviously a matter of opinion. Unless it is proven otherwise, however, the development of structural surfaces some distance above stream level must be considered a possible explanation.

According to Trowbridge (1935, p. 62, 75) the existence of two peneplanes is further indicated by the relation of various parts of the upland to certain upland gravels and to Nebraskan glacial drift. The basis of this conclusion is as follows: Upland gravels belonging to the Windrow

formation (Thwaites and Twenhofel, 1920, p. 293-314) are found only on the higher parts of the upland which have been interpreted by both Trowbridge and the writer as peneplane remnants. Trowbridge believes that the gravels were deposited after the peneplane was developed, thereby dating the upper part of the upland as pre-gravel (pre-late Tertiary according to Trowbridge) in age. Nebraskan drift, on the other hand, is found west of the Mississippi River on both the upper and lower parts of the upland, but never in the valleys cut below the upland. Trowbridge has interpreted this fact to indicate that the drift was deposited before the present valleys were cut. Accordingly he concludes that the present topography consists of (1) an upper peneplane which was developed prior to the deposition of the upland gravels (pre-late Tertiary), (2) a lower peneplane which was developed, but undissected, at the time of Nebraskan glaciation, and (3) valleys cut below the upland in post-Nebraskan time.

The writer has, on the basis of other evidence already presented, interpreted the areas on which the gravels are found as peneplane remnants, thus giving additional support to the first point in Trowbridge's interpretation. The validity of the second and third points of the argument rests on the relations of the lower upland surface to Nebraskan glacial drift. At the present time published information on the distribution of the older glacial drift along the western side of the Driftless Area is extremely meager. This situation will, however, be materially modified in the near future by the publication of a paper by A. C. Trowbridge (personal communication) giving a detailed discussion of the older drift including its bearing on the erosional history of the region. In anticipation of this discussion it may be said that proof of the suggested hypothesis depends on whether the absence of Nebraskan glacial drift from these valleys necessarily indicates that the upland was undissected at the time the drift was deposited. As an alternative explanation it is suggested that drift deposited in the valleys may have been removed by further dissection.

If we accept the suggestion of Trowbridge (1934, p. 524; 1935, p. 75-76) that the major stream of the region—the Mississippi—was displaced to its present course by the eastward advance of the Nebraskan glacier, rapid intrenchment of the main and tributary streams following the displacement seems a likely possibility. This would probably accelerate the removal of all drift from the valleys. Consequently, while there is no evidence in the Driftless Area necessitating the above conclusions, there is nothing incompatible with such an interpretation.

Thwaites' hypothesis of single peneplane.—Thwaites (1935, p. 120) has recently recognized

"one mature surface with a relief of about 200 feet, that is essentially Trowbridge's Lancaster peneplane."

The expression, a "mature surface" normally suggests one in which the relief is at a maximum and, hence, quite different from the usual conception of a peneplane. The description, however, seems to include as peneplane remnants the lower parts of the upland which are here interpreted, from evidence in the Driftless Area, as stripped structural surfaces, and to regard as monadnocks the higher areas along the crests of the cuestas and a few divides which constitute Trowbridge's Dodgeville peneplane and which are interpreted in this paper as peneplane remnants. The writer has already set forth his reasons for believing that the highest beveled surfaces are peneplane remnants (Dodgeville), and the lower benches stripped structural surfaces.

Alternative hypothesis of single peneplane.—On the basis of observations made in the Driftless Area and set forth in this paper the writer has already in effect presented his own interpretation—namely, that the upland of the Driftless Area contains remnants of but one peneplane which is the Dodgeville of Trowbridge (Fig. 2D). These remnants, as shown in Figure 4, are found along the crests of the cuestas developed on the Prairie du Chien, the Platteville-Galena, and the Niagaran, and along some of the principal divides, notably that between the Kickapoo and Mississippi drainage basins. No peneplane remnants are found on the ragged cuesta developed on a resistant stratum within the St. Croixan in the area covered by the profiles. Since this cuesta does not come up to the level of the others, none is to be expected. Those parts of the upland not included in the peneplane remnants shown in Figure 4 are considered stripped structural surfaces, but the possibility that these surfaces were formed in an incomplete erosion cycle earlier than the present cannot be definitely eliminated.

CONCLUSION

As observed on the projected profiles the higher parts of the upland surface along the crests of the cuestas and the principal divides bevel the structure. Since each of these higher areas can be projected to similar adjoining areas to form a surface of low relief they are taken as conclusive evidence of a single peneplane. Though Trowbridge's work outside of the Driftless Area may necessitate recognition of a second peneplane, all available evidence observed in the Driftless Area permits the interpretation that the stripped structural surfaces, together with the stream valleys cut below the upland, were formed in the present cycle following rejuvenation of the Dodgeville peneplane.

RELATION OF DODGEVILLE PENEPLANE TO PENEPLANE OF CENTRAL AND NORTHERN WISCONSIN

GENERAL STATEMENT

The existence of a strikingly even upland in the pre-Cambrian crystalline area of central and northern Wisconsin has long attracted the attention of geologists. As the existence of the Dodgeville peneplane has been established in southwestern Wisconsin, it is of interest to determine, if possible, the relation of this peneplane to the surface which bevels the crystallines farther north.

POSSIBLE INTERPRETATIONS

The earliest discussion of the erosion surface in the crystalline area was by Van Hise (1896, p. 57-59). It was his opinion, arrived at without the aid of accurate elevations or topographic maps, that this surface in north central Wisconsin continues to the south on the tops of the buttes of St. Croixan sandstone. Although he did not carry the surface to the dolomite cuestas on which the Dodgeville peneplane is developed in the southwestern part of the state, he at least had the idea of a peneplane which beveled both the crystallines and the sediments to the south. Van Hise also recognized the existence of another and much older peneplane beveling the pre-Cambrian formations where they are exposed by partial removal of Cambrian sediments. Weidman (1903, p. 289-313), working in central and northern Wisconsin, concluded that the peneplane on the crystallines could be traced southward to the place where it passed beneath the Cambrian sediments. Subsequent writers (Martin, 1911, p. 116; 1932, p. 369; Thwaites, 1931, p. 743-744), almost without exception, have favored Weidman's conclusions. Recently, however, Douglas Johnson has reopened the question as to the possibility of a Tertiary peneplane beveling both the crystallines and the sediments to the south.

Any interpretation of the relationship must take into consideration: (1) the southward sloping Dodgeville peneplane, and (2) the more steeply dipping pre-Cambrian surface. Given these conditions, the peneplane surface beveling the crystallines may be interpreted as:

(1) the northward extension of the Dodgeville peneplane beveling the crystallines of the central and northern parts of the state as well as the sediments of southwestern Wisconsin, as suggested by Van Hise (1896, p. 57-59) and Johnson (personal communication) (Fig. 6A),

(2) a pre-Cambrian peneplane resurrected by pre-Dodgeville erosion (Fig. 6B),

(3) a pre-Cambrian peneplane resurrected by post-Dodgeville erosion (Fig. 6C),

(4) a combination of 1 and 3 (Fig. 6D), or
(5) a combination of 2 and 3 (Fig. 6E).

There is nothing in Weidman's paper to indicate whether he considered 2, 3, or 5 the correct interpretation.

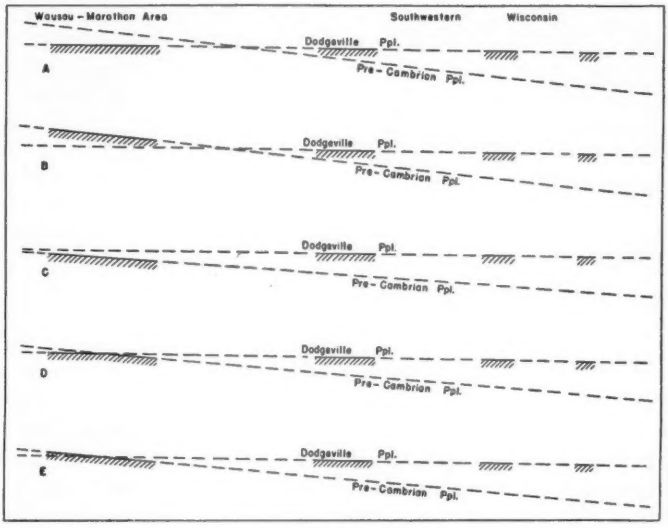


FIGURE 6.—*Possible relations of Dodgeville and northern Wisconsin peneplanes*

Assuming no warping, which might have complicated the situation in a variety of ways.

EVIDENCE FROM PROJECTED PROFILES

To aid in eliminating as many of these alternative hypotheses as possible, profiles were constructed of the Wausau and Marathon quadrangles, which are the only topographic maps available in the crystalline area of the State (Pl. 3, figs. 1, 2). As seen from Plate 3, figure 2 the Wausau-Marathon area has the appearance of a moderately undulating surface surmounted by a few very prominent peaks, the highest one being Rib Hill. Closer examination reveals considerable irregularity, particularly in the central part of the area. The western part of the area apparently slopes gently southward with few irregularities. The east central part is marked by a low east-west ridge. North of the ridge the surface is almost horizontal, and south of the ridge the upland breaks away irregularly to the valley of the Wisconsin River intrenched some distance below.

Attempts to correlate these surface irregularities with differences of

underlying rock have been only partially successful. The outcrop of Huronian quartzite is consistently marked by striking topographic elevations. The three prominent peaks shown in Plate 3 near the center of the area are of this formation. No other correlation was found between the smaller irregularities of the surface and the underlying rocks.

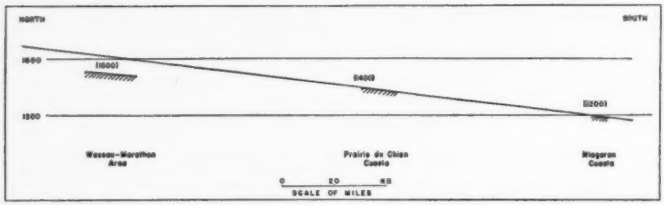


FIGURE 7.—*Relation of Wausau-Marathon area to Dodgeville peneplane projected*

Wausau-Marathon area is somewhat below Dodgeville peneplane as projected from crests of Prairie du Chien and Niagaran cuestas.

It is evident from photographs of the profiles (Pl. 3, figs. 1, 2) and relations indicated that the ideally developed even surface described by most writers disappears with more careful examination, as has been emphasized by Thwaites (1931, p. 744). In fact, it is exceedingly difficult from the small area which it was possible to profile to know which part is typical of the surrounding region. Considering only the bench in the northeastern part of the area, one finds it to be slightly below the northward projection of a line touching the remnants of the Dodgeville peneplane on the crests of the Prairie du Chien and Niagaran cuestas. The bench is not sufficiently below the projection of this line to eliminate the possibility that it is a remnant of a somewhat warped Dodgeville peneplane (Fig. 7). If, on the other hand, one considers the more even slope in the western part of the area, it is found to be not far out of line with the northward projection of the pre-Cambrian-Cambrian unconformity.

Since outliers of Cambrian sandstone would be removed wherever the pre-Cambrian is beveled by a more gently dipping post-Cambrian peneplane, the report of Thwaites [3] that he has observed sandstone float near Wausau suggests that the Dodgeville beveling of the pre-Cambrian is limited to the area north of Wausau. Although the available data will permit a beveling of the pre-Cambrian surface north of the Wausau-Marathon area by a post-Cambrian peneplane, warping of the pre-Cambrian peneplane will also account for the conditions observed.

[3] Letter to Douglas Johnson. December 9, 1932.

CONCLUSIONS

The principal conclusion to be drawn from this part of the investigation is that the available facts are not sufficient to permit a satisfactory solution of the problem of peneplane relationships. Certain positive conclusions are nevertheless justified. All available evidence seems to eliminate the possibilities that the Wausau-Marathon area is either wholly or in part a pre-Cambrian peneplane resurrected by pre-Dodgeville erosion. From the data at hand, there is apparently no means of eliminating any of the remaining hypotheses. The close approximation in this area of the northward projections of the Dodgeville and pre-Cambrian peneplanes is taken to indicate that these two do intersect. In this case their line of intersection, if not within this area, is a relatively short distance either to the north or south.

Additional field information is necessary for a satisfactory solution of the problem. In this connection the accurate location of sandstone outliers in the northern part of Wisconsin would be of particular significance.

DEVELOPMENT OF THE KICKAPOO VALLEY

GENERAL STATEMENT

Having considered the development of the upland into which the Kickapoo Valley has been cut, we may now turn to the valley itself. Development of the valley will be considered in two sections. The first concerns the underfit character of the stream, together with the related presence of extensive rock terraces in the upper part of the valley. The second has to do with deposits and forms resulting from aggradation of the valley.

UNDERFIT CHARACTER OF KICKAPOO RIVER

The Kickapoo River, throughout its length of about 85 miles, is intrenched 300 to 400 feet below the upland. When studied in detail the appearance of the valley varies considerably. In the lower 55 miles, below LaFarge, the stream occupies a valley that has been opened out from a quarter of a mile to half a mile in width (Fig. 8). North of LaFarge for 12 miles (18 miles along the river) the stream is intrenched in rock terraces in the bottom of a wider valley (Fig. 9). In the headwater portion of the stream the valley is much narrower and without unusual features. Throughout its entire length the present valley floor is underlain by alluvial material, the river having been raised to its present level (50 to 125 feet above bedrock) as a result of aggradation. It is on this valley floor that the Kickapoo River has developed numerous meanders. In contrast to the small meanders of the river, the valley is marked by several large symmetrical bends (two of them cut-off loops) which have approximately identical radii of curvature appropriate to the meanders

of a large stream (Fig. 8). The small meander pattern of the stream itself is developed not only on the open valley floor of the southern part of the stream but also in the headwater portions where the small meanders

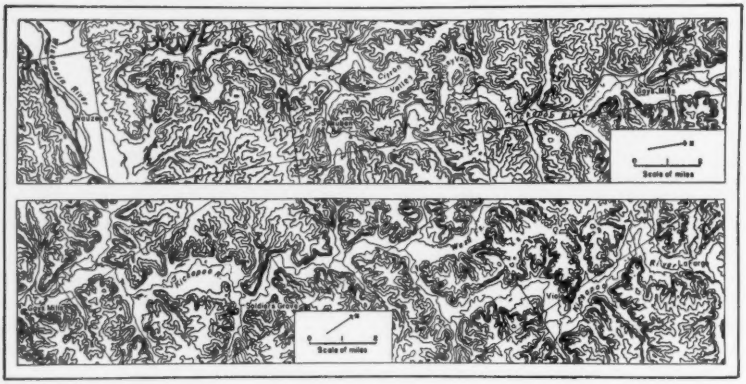


FIGURE 8.—*Maps of lower (above) and middle sections of Kickapoo valley*
Contour interval 100 feet. Topography from U. S. Geol. Survey. For location of maps *see* insert of Figure 9.

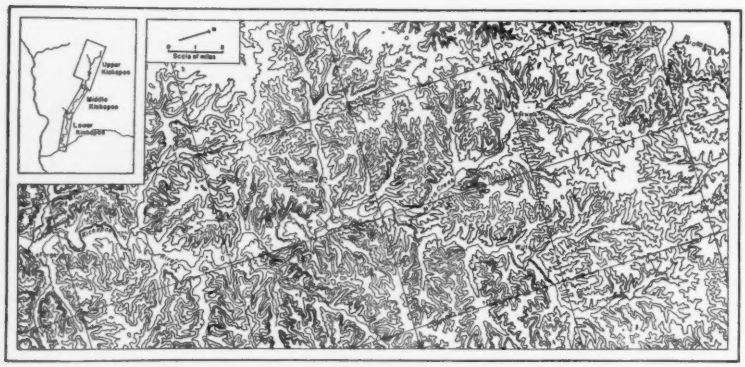


FIGURE 9.—*Upper Kickapoo valley*
Contour interval 100 feet. Topography from U. S. Geol. Survey.

are incised below rock terraces (Fig. 9). A stream exhibiting such a contrast between large meanders of the valley and small meanders of the stream is commonly termed an "underfit stream." The concept of underfit streams is based on the general observation that large streams produce meanders of larger radius of curvature than do small streams. The exist-

ence of streams having small meanders in valleys possessing a large meander pattern has, therefore, usually been taken as *prima facie* evidence of a decrease in discharge.

In order to understand misfit streams and interpret the features of the Kickapoo Valley, the factors involved in meander development will be critically examined to determine the extent to which meander size is dependent on stream size, and to determine whether other factors are in part responsible for the misfit relations.

MEANDERS DEVELOPED ON FLOODPLAINS

It is a frequent characteristic of streams flowing on a floodplain to follow more or less regularly developed serpentine curves, known as meanders. Successive surveys of large streams, such as the Mississippi, show that these meanders grow outward and migrate down valley. Furthermore, meanders are frequently cut off by the stream's taking a shorter course across the necks of former meanders. Such cut-offs may be formed either by intercision of the stream on the two sides of the meander neck or by the stream's flowing across the neck of the meander in time of flood and maintaining the new course when the flood recedes. The latter type has been termed by Fisher (1906, p. 15-16) the short-cut process. The newly formed curve which results from the cut-off immediately begins to grow outward and down valley developing into a new meander and repeating the evolution.

Mechanics of meander development.—Several factors tend to produce more intensive erosion on one bank of a stream than on the other.[4] These may be considered in two groups, the physical and the geological. (It is recognized that the geological factors are, in the last analysis, largely physical.) The physical factors include centrifugal force exerted by water moving in a curved path, deflection due to Coriolis forces,[5] and tidal attractions. The geological factors include such things as variation of rock types on the two sides of a stream and structural relations which favor unequal erosion along the two banks. In this discussion the geological factors are assumed to be nonexistent except when specifically stated. Although the deflection due to Coriolis forces may be of significance in high latitudes and the tidal effects appreciable in some of the larger streams, the dominant factor tending to produce unequal erosion on the two banks of a stream is centrifugal force.

[4] The following discussion has drawn freely on the works of Rudski (1911) and Exner (1931).

[5] Without reference to or apparent knowledge of the previous work of Coriolis, Ferrel (1859, p. 307) summarized the forces arising from the earth's rotation—a special case of Coriolis forces—as follows: "in whatever direction a body moves on the surface of the earth, there is a force arising from the earth's rotation, which deflects it to the right in the northern hemisphere, but to the left in the southern." This is the statement commonly referred to in geological and geographical literature as "Ferrel's Law."

Any mass moving in a curved path exerts a centrifugal force. In a meander the force is applied to the outside bank of the stream (Fig. 10). Because of the centrifugal force the dynamic axis (line of greatest velocity) does not follow the center of the stream but is deflected toward the outside of the curve. The dynamic axis is near the surface, though usually not on it, and since the faster parts of the stream near the dynamic axis are more extensively deflected, a cross-circulation or screw-motion is set up. Thus the water, in addition to its downstream flow, moves toward the outer bank near the surface, downward along the outer bank, and toward the convex bank along the bottom. The deflection of the dynamic axis results in more extensive erosion along the outer bank, and together with the cross-circulation thus set up is responsible for the movement of eroded material toward the convex bank and for the asymmetrical transverse channel profile across the meander. Leighly (1932, 1934) presents data supporting the theory that, because of the asymmetrical transverse profile, erosion along the outer bank of the stream is further increased by the greater turbulence which develops near the base of the steeper side of the channel. Accordingly, the corrasive action of the stream is greater along that part of the curve in which the dynamic axis is nearer one bank of the stream, as is shown in Figure 10, or, in other words, to the outside and downstream side of the curve. This explains the lateral and downstream migration of the meanders observed in the surveys of the Mississippi River and other streams.

It remains to be determined, however, whether the growth continues indefinitely until a cut-off occurs, or, if not, what factors limit it. Further consideration of centrifugal force goes far to answer this question.

Centrifugal force, f, the outwardly directed reaction to the inwardly directed centripetal force, is a function of the mass, m, the velocity, v, and the radius of curvature, r, as expressed in the following equation:

$$f = \frac{mv^2}{r} \tag{1}$$

By dividing both sides of the equation by m, we obtain the centripetal force per unit of mass, or, in other words, the centripetal acceleration, a,

$$\frac{f}{m} = \frac{v^2}{r} = a \tag{2}$$

The problem is to investigate the behavior of a when applied to meander growth. The velocity is assumed to be proportional to the force accelerating the stream—namely, the effective component of gravity, thus

$$v = kg \sin i \tag{3}$$

where g is the acceleration due to gravity, i the angle of slope of the stream, and k the constant of proportionality. This assumed proportionality is strictly true only when the flow is nonturbulent. It is, however, sufficiently near correct for purposes of this discussion.

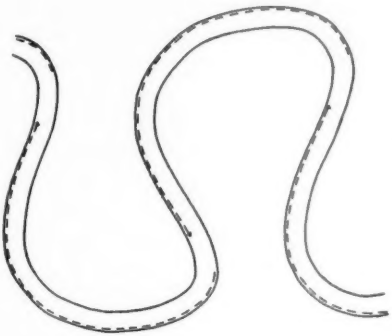


FIGURE 10.—*Sections of greatest erosion on a meander*
Broken line drawn near bank subjected to more intense erosion.

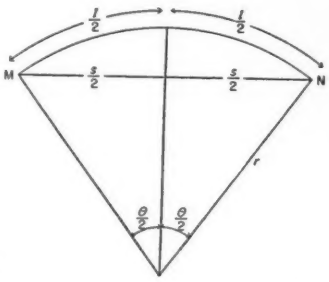


FIGURE 11.—*Elements of meander growth*

For meandering streams the angle i is small. Therefore we may set

$$sin\ i = i = tan\ i = \frac{h}{l} \tag{4}$$

where i is measured in radians and $\frac{h}{l}$ is the stream's gradient; accordingly

$$v = \frac{kgh}{l} \tag{5}$$

and

$$a = k^2g^2h^2 \cdot \frac{1}{l^2r} \tag{6}$$

Let us now apply these relations to a stream passing through two points, M and N, a distance s apart and having a difference of elevation h. The stream segment MN is represented as an arc of a circle having a length l, a radius of curvature r, and subtending an angle θ (Fig. 11). Let the stream by the process of meander development be imagined to occupy all possible positions from a straight line between M and N to an arc of very large circumference, subject only to the limitations that it

must remain a segment of a circle, and must pass through points M and N of fixed distance s apart, and of fixed difference of elevation h.

As the meander of the stream undergoes this development the angle θ subtended by the arc will change from a mathematical limit of zero when the arc approaches the straight line MN to 2π radians as the length of l becomes large with respect to the distance s between the two fixed points. While the length of the arc, l, increases throughout the meander development, the radius of curvature, r, decreases for a time, reaches a minimum, and thereafter increases.

Taking θ as the independent variable, we will examine the variation of the centripetal acceleration a as a function of θ. From Figure 11

$$r = \frac{s}{2 \sin \frac{\theta}{2}} \tag{7}$$

and

$$\frac{l}{2} = r\frac{\theta}{2} = \frac{\theta}{2} \cdot \frac{s}{2 \sin \frac{\theta}{2}} \tag{8}$$

or

$$l = \frac{\theta s}{2 \sin \frac{\theta}{2}} \tag{9}$$

Introducing the values for r and l of equations (7) and (9) into equation (6), we obtain

$$a = k^2g^2h^2 \cdot \frac{4 \sin^2 \frac{\theta}{2}}{\theta^2 s^2} \cdot \frac{2 \sin \frac{\theta}{2}}{s} = \frac{k^2g^2h^2}{s^3} \cdot \frac{\left(2 \sin \frac{\theta}{2}\right)^3}{\theta^2} \tag{10}$$

from which the acceleration a is readily determined for all values of θ between the extremes of zero and 2π radians. The value of a may be expressed in multiples of the constant factor $\frac{k^2g^2h^2}{s^3}$ of equation (10).

The factor $\frac{\left(2 \sin \frac{\theta}{2}\right)^3}{\theta^2}$ approaches zero as a limit as θ approaches zero or 2π radians. It has a maximum value when θ equals 2 radians (114° 35.5′).

It is also useful to know the ratio of $\frac{r}{s}$ and $\frac{l}{s}$ for the various values of

θ. These are given directly by equations (7) and (9):

$$\frac{r}{s} = \frac{1}{2 \sin \frac{\theta}{2}}$$

and

$$\frac{l}{s} = \frac{\theta}{2 \sin \frac{\theta}{2}}$$

We can now evaluate the acceleration a in terms of the constant factor, $\frac{k^2g^2h^2}{s^3}$, and the length of the meander arc, l, and the radius of curvature, r, in terms of the length s, for various values of θ from zero to 2π. This has been done and the values are given in Table 1. The same results are shown graphically in Figure 12.

TABLE 1.—*Variation of centripetal acceleration, radius of curvature, and length of arc during meander growth*

Angle subtended by meander (radians)	Centripetal acceleration $\frac{k^2g^2h^2}{s^3}$	Radius of curvature (s)	Length of arc (s)
0	0	∞	1.00
$\pi/6$	0.505	1.930	1.01
$2\pi/6$	0.912	1.000	1.05
$3\pi/6$	1.14	0.707	1.11
$4\pi/6$	1.18	0.578	1.21
$5\pi/6$	1.05	0.518	1.35
π	0.810	0.500	1.57
$7\pi/6$	0.537	0.518	1.84
$8\pi/6$	0.294	0.578	2.42
$9\pi/6$	0.127	0.707	3.33
$10\pi/6$	0.0365	1.000	5.24
$11\pi/6$	0.00418	1.930	11.1
$12\pi/6$	0	∞	∞

It can thus be seen that the centrifugal force, which is apparently so active in developing meanders, varies greatly in intensity from the early stages of development to the later ones. As the force decreases, the deflection of the dynamic axis is lessened. This, in turn, reduces the greater erosion on the outer bank of the stream and the asymmetry of the transverse profile, thereby reducing whatever part of meander growth results from increased turbulence along the steeper bank of the stream. Consequently, it must be recognized that although it is true that a meander grows until it is cut off, that is only part of the story. To get a complete picture of the phenomenon it should be noted that unless the cut-off occurs in the early stages of meander growth, the rate of growth will be

effectively reduced by a decrease in centrifugal force. Furthermore, when meander growth starts from a cut-off, the first part of the develop-

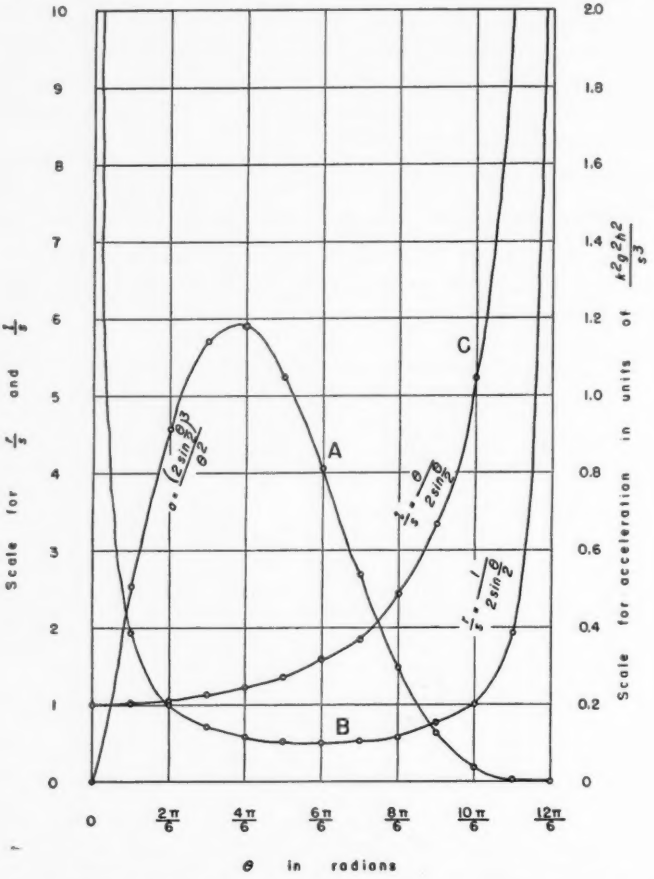


FIGURE 12.—*Variation of centrifugal acceleration a, (curve A), radius of curvature r, (curve B) and length of arc l, (curve C) during meander growth*

ment—that characterized by a decreasing radius—is to a large extent omitted. Although the absolute limit of meander growth under any given conditions cannot be predicted as yet, the rate of growth decreases rapidly throughout most of the period of development.

The size of meanders is commonly expressed in terms of width of meander belt. Because there is some variation in the use of the term "meander belt" it is defined for purposes of this discussion as the area between lines drawn on each side of the stream tangent to the outside of successive meanders. Thus it becomes wider as the meanders grow and collapses locally to the width of the stream when a cut-off occurs. It is, therefore, only the greatest or limiting width of the meander belt that is diagnostic in the case of any particular stream.

Relation of meander size to stream size.—Having considered the causes and limits of meander growth, we are now in a position to return to the original question of determining the connection between stream size—or better stream discharge—and the size of meanders. The centripetal acceleration applies to each unit of mass in the stream, so that, other factors remaining constant, the total centripetal force exerted on a stream (or the opposite centrifugal force exerted by a stream) at any point would be derived to a first approximation from the general expression

$$F = \frac{mv^2}{r} \tag{11}$$

in which m is the mass, v the velocity, and r the radius of curvature. Considering a unit length of the stream, the mass directed against the outside bank of the meander is essentially proportional to the cross-sectional area of the channel at that point. The discharge of the stream is the product of its cross-sectional area and its velocity. Making these substitutions equation (11) becomes

$$F = \frac{Av^2}{r} = \frac{Dv}{r} \tag{12}$$

in which A denotes the cross-sectional area, and D the discharge of the stream. Thus, if the rate of meander growth is essentially proportional to the centrifugal force, as is suggested in this paper, the rate of meander growth, and, therefore, the size attained in a given period of meander development, would also be related to the discharge of the stream.

To check the suggestions regarding relation of meander size to stream size, data on the maximum or limiting width of meander belts have been assembled from topographic maps. In presenting some data on this point Jefferson (1902, p. 380) concluded that the average maximum ratio of width of meander belt to width of stream on 23 American and European rivers flowing on floodplains was 17.6:1. Jefferson discarded measurements on several streams in which the ratio of the width of meander

belt to the distance from meander to meander along the river axis was considered too small. The reason for this rejection is not clear but presumably it was because cut-offs could not take place as readily where the meanders were more widely spaced along the stream. In view of the fact that a cut-off will take place in time even where there is only one meander in the stream, such restriction of data seems unnecessary.

For additional information the writer has examined all available topographic sheets of the United States Geological Survey. The streams studied include only those with well developed meanders and sufficient size to be mapped by more than a single line. It is considered permissible to include in the data such cut-off meanders or ox-bows as may still be connected with the stream by smooth meander-like curves. The absolute maximum width of the meander belt in a given part of a stream may conceivably be abnormal due to special conditions. The normal limiting width of a meander belt is, therefore, taken as the maximum width consistently developed by the regular meanders of the stream. The width of the meander belt given in the table is the normal limiting width.

In the measurements made by Jefferson and by the writer, stream width has been used as a measure of stream size or discharge on the assumption that width is proportional to discharge. Such an assumption is subject to two chief sources of error: (1) the cross-sections of all streams flowing on floodplains may not be essentially similar, and (2) all streams on floodplains have not the same velocity. In the absence of satisfactory data on the first point it may be pointed out that the configuration of channels of streams flowing on floodplains is developed to a large extent by the streams themselves, in an effort to acquire that cross-section which will transmit the flow with minimum resistance. There is, therefore, some basis for assuming a reasonable uniformity of channel cross-section. Velocity of meandering streams is known to vary to some extent, and error is thus introduced into the problem; it is believed, however, that where a large number of streams are involved the net error will not be great. Because of the difficulty of obtaining data on channel form and discharge for a great number of meandering streams, it has seemed advisable to use stream width as an index of stream discharge in spite of the errors involved. Such errors are probably no greater than those involved in measuring stream widths on topographic maps.

The data obtained by the writer are summarized in Table 2 and in graphic form in Figure 13. Since the distribution of points on the graph suggests a straight line relationship, the line of mean ratio was computed by the least square method, and found to be 16:1 for streams 100 feet wide, 12:1 for streams 1000 feet wide, and 11:1 for streams 3000 feet wide.

TABLE 2.—*Meander Ratios of Streams on Floodplains*

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
Animas R.	100	1,300	13	Durango, Colo.	1:62,500
Arkansas R.	1,000	11,000	11	Gillett, Ark.	1:62,500
Arkansas R.	400	5,000	12	Kaw, Okla., Kan.	1:62,500
Bear R.	180	2,200	12	Randolph, Utah, Wyo.	1:125,000
Bear R.	200	1,800	9	Logan, Utah, Wyo.	1:125,000
Beaver Cr.	90	1,200	13	Bowdoin, Mont.	1:62,500
Belle Fourche R.	180	4,600	25	Newell, S. D.	1:125,000
Big Blue R.	600	4,000	7	Junction City, Kansas	1:125,000
Big Muddy R.	225	4,500	20	Alto Pass, Ill.	1:62,500
Big Muddy Cr.	100	1,300	13	Homestead, Mont.	1:62,500
Big Sioux R.	175	2,600	15	Elk Point, S. D., Neb., Iowa	1:125,000
Black R.	130	1,500	11	Reyno, Ark, Mo.	1:62,500
Black R.	250	2,500	10	Lowville, N. Y.	1:62,500
Black R.	330	3,300	10	North Bend, Wis.	1:62,500
Blackfoot R.	50	600	12	Lanes Creek, Idaho	1:62,500
Black Warrior R.	260	3,800	15	Eutaw, Alabama	1:62,500
Brazos R.	150	3,500	23	Courtney and Howth, Texas	1:31,680
Canadian R.					
North Fork	100	1,400	14	McLoud, Okla.	1:62,500
North Fork	400	6,400	16	Canadian, Okla.	1:125,000
Chariton R.	60	1,400	14	Queen City, Mo.	1:62,500
Cherry Valley Cr.	90	1,000	11	Cooperstown, N. Y.	1:62,500
Cimarron R.	550	6,000	11	Perkins, Okla.	1:62,500
Coldwater R.	90	2,500	30	Crenshaw, Miss.	1:62,500
Connecticut R.	400	3,500	9	Woodsville, N. H., Vermont	1:62,500
Connecticut R.	75	700	9	Indian Stream, N. H., Vermont	1:62,500
Crooked Cr.	50	1,200	24	Carlyle, Ill.	1:62,500
Current R.	200	2,300	11	Reyno, Ark.	1:62,500
Cuyahoga R.	110	900	8	Cleveland, Ohio	1:62,500
Dead R.	100	1,400	14	Stratton, Me.	1:62,500
Des Moines R.	220	2,800	13	Des Moines and Waukee, Iowa	1:62,500
Des Moines R.	650	4,500	7	Milo and Mitchellville, Iowa	1:62,500
Duchesne R.	200	2,100	10	Vernal, Utah	1:125,000
Edisto R.	230	1,600	7	Lodge, S. C.	1:62,500
Elkhorn R.	150	3,000	20	Fremont, Neb.	1:125,000
Embarrass R.	125	1,400	11	Vincennes, Ind., Ill.	1:62,500
Feather R.	225	4,000	18	Knight's Landing, Cal.	1:31,680
Feather R.	225	3,500	15	Gridley and Honcut, Cal.	1:31,680
Genesee R.	125	1,500	12	Caledonia, N. Y.	1:62,500
Grand R.	130	1,300	10	Lansing, Mich.	1:62,500
Grand R.	300	4,600	15	Marshall, Mo.	1:125,000
Grand R.	400	3,500	9	Hale, Mo.	1:62,500
Grand R.	215	2,800	13	Chillicothe, Mo.	1:62,500
Green R.	100	1,200	12	Gros Ventre, Wyo.	1:125,000
Housatonic R.	100	1,100	11	Sheffield, Mass., Conn., N. Y.	1:62,500

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
Iowa R.	160	2,200	14	Amana, Iowa	1:62,500
James R.	200	2,400	12	Lamoure, N. D.	1:125,000
Kaskaskia R.	100	1,600	16	Carlyle, Ill.	1:62,500
Kickapoo R.	80	900	11	Wauzeka, Wis.	1:62,500
Kootenai R.	450	4,800	11	Priest Lake, Idaho	1:250,000
Lamoille R.	150	1,800	12	Mount Mansfield, Vermont	1:62,500
Licking R.	90	2,000	22	Salt Lick, Ky.	1:62,500
Little R.	210	2,300	11	Pollack, La.	1:62,500
Little R.	60	1,800	30	Buckholts, Tex.	1:31,680
Little Big Horn	120	1,400	12	Custer Battlefield, Montana	1:24,000
Little Wabash R.	190	4,500	24	New Haven, Ill., Ind., Ky.	1:62,500
Medicine Bow R.	100	1,800	18	Saddleback Hills, Wyo.	1:62,500
Meherrin R.	100	1,300	13	Arringdale, Va., N. C.	1:62,500
Merced R.	100	1,250	12	Turlock, Calif.	1:31,680
Merrimack R.	400	3,100	8	Penacook, N. H.	1:62,500
Milk R.	175	3,500	20	Nashua, Mont.	1:62,500
Milk R.	200	1,800	9	Assiniboine, Mont.	1:62,500
Milwaukee R.	150	2,000	13	West Bend, Wis.	1:62,500
Mississippi R.	2,400	31,000	13	Baton Rouge, La.	1:62,500
Mississippi R.	2,600	35,500	14	Natchez, La., Miss.	1:62,500
Mississippi R.	2,900	30,000	10	Deer Park, La., Miss.	1:62,500
Mississippi R.	3,000	36,000	12	Refuge, Ark., Miss.	1:62,500
Mississippi R.	200	2,400	12	Cuyuna, Minn.	1:62,500
Missouri R.	2,500	18,400	7	Marshall, Mo.	1:125,000
Missouri R.	1,400	20,000	14	St. Joseph, Mo.	1:62,500
Missouri R.	1,100	14,400	13	Bismarck, N. D.	1:125,000
Missouri R.	1,800	18,000	10	Chelsea, Mont.	1:62,500
Missouri R.	1,250	14,000	11	Frazer, Mont.	1:62,500
Muddy R.	90	1,800	20	Herrin, Ill.	1:62,500
Muskingum R.	400	4,500	11	Conesville, O.	1:62,500
Nemadji R.	130	1,100	8	Superior, Wis., Minn.	1:62,500
Nemaha R.	100	1,800	18	Falls City, Neb.	1:62,500
Oconee R.	300	1,700	6	Irwinton, Ga.	1:62,500
Ohio R.	2,500	34,000	14	Henderson, Ky., Ind.	1:62,500
Ouachita R.	500	3,600	7	Camden, Ark.	1:125,000
Passaic R.	150	1,400	9	Morristown, N. J.	1:62,500
Patoka R.	90	1,600	18	Petersburg, Ind.	1:62,500
Pawpaw R.	50	800	16	Hartford, Mich.	1:62,500
Payette River North Fork	175	2,000	11	Meadows, Idaho	1:125,000
Pearl R.	225	2,000	9	Jackson, Miss.	1:62,500
Pecatonica R.	130	1,500	11	Pecatonica, Ill., Wis.	1:62,500
Pecatonica R.	100	1,500	15	South Wayne, Wis., Ill.	1:62,500
Platte R.	160	2,700	17	Leavenworth, Kan., Mo.	1:62,500
Raccoon R.	200	2,500	12	Des Moines and Waukee, Iowa	1:62,500

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
Raisin R.	250	2,800	11	Dundee, Mich.	1:62,500
Red R.*	250	14,000	56	Gainesville and Denison, Tex, Okla.	1:125,000
Republican R.	400	5,000	12	Clay Center, Kan.	1:125,000
Republican R.	250	7,200	29	Holdrege, Neb., Kan.	1:125,000
Rio Grande	350	5,500	16	Mission and San Juan, Texas	1:62,500
Rock R.	250	2,400	10	Waterloo, Wis.	1:62,500
Rough R.	100	1,500	15	Whitesville, Ky.	1:62,500
Sacramento R.	200	3,500	17	Kirkville, Calif.	1:31,680
Sacramento R.	320	4,500	14	Compton Landing, Calif.	1:31,680
Sacramento R.	500	6,000	12	Chico Landing, Calif.	1:31,680
St. Francis R.	200	4,500	22	Princedale, Ark.	1:62,500
St. Francis R.	80	2,000	25	Valley Ridge, Mo.	1:62,500
St. Francis R.	115	1,800	16	Puxico, Mo.	1:62,500
St. Mary's R.	90	1,000	11	Celina, Ohio.	1:62,500
St. Mary's R.	75	1,000	13	Moniac, Ga., Fla.	1:62,500
Sangamon R.	150	1,700	11	Springfield, Ill.	1:62,500
South Fork	60	1,100	18	Divernon, Ill.	1:62,500
San Joaquin R.	200	3,500	17	Vernalis and Lathrop, Calif.	1:31,680
San Joaquin R.	150	2,200	15	Mitchell School and Crows Landing, Calif.	1:31,680
San Joaquin R.	150	1,500	10	Turner Ranch and San Luis Ranch, Calif.	1:31,680
San Joaquin R.	250	3,500	14	Mendota, Calif.	1:31,680
Santee R.	320	4,200	13	Chicora and Manning, S. C.	1:62,500
Savannah R.	190	3,000	16	Shirley, Ga., S. C.	1:62,500
Savannah R.	165	2,000	12	Peeples and Ellenton, Ga., S. C.	1:62,500
Savannah R.	250	4,000	16	Augusta, Ga., S. C.	1:62,500
Scioto R.	380	6,500	17	Chillicothe and Roxabell, Ohio	1:62,500
Smoky Hill R.	200	4,800	24	Russell, Kansas	1:125,000
Solomon R.	200	3,600	18	Beloit, Kansas	1:125,000
Spoon River	120	2,000	17	Avon, Illinois	1:62,500
Snoqualamie R.	200	2,200	11	Sultan, Wash.	1:125,000
Stanislaus R.	50	1,000	20	Ripon and Salida, Calif.	1:31,680
Sulphur R.	115	2,300	20	Texarkana, Tex., Ark.	1:62,500
Sunflower R.	150	1,800	12	Cleveland, Miss.	1:62,500
Thompson R.	190	2,000	10	Chillicothe, Mo.	1:62,500
Trempealeau R.	90	1,000	11	Fountain City, Wis., Minn.	1:62,500
Trinity R.	100	1,100	11	Grapevine, Tex.	1:62,500
Tualatin R.	150	1,900	13	Tualatin, Oregon	1:62,500
Tuscarawas R.	350	3,000	8	Newcomerstown and Coshocton, Ohio	1:62,500
Tygarts R.	80	1,400	17	Greenup, Ky., Ohio.	1:62,500
Verdigris R.	160	2,800	17	Claremore and Muskogee, Okla.	1:125,000
Wabash R.	1,000	17,000	17	New Haven, Ill., Ind., Ky.	1:62,500
Wabash R.	900	8,500	9	Mount Carmel and New Harmony, Ind., Ill.	1:62,500
Wabash R.	700	6,500	9	Vincennes, Ind., Ill.	1:62,500
Wabash R.	300	4,000	13	Birds, Ill., Ind.	1:62,500
Wallkill R.	90	800	9	Goshen, N. Y.	1:62,500

* Width of stream probably less than that responsible for meander belt.

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
Wapsipinicon R.	175	2,500	14	DeWitt, Iowa	1:62,500
Washita R.	100	1,600	16	Tishomingo, Okla.	1:125,000
White R.	425	6,500	15	Augusta, Arkansas	1:62,500
White R.	90	1,300	14	Meeker, Colorado	1:62,500
Wichita R.*	110	4,250	39	Iowa Park, Texas	1:31,680
Wills Cr.	60	1,200	20	Cambridge, Ohio	1:62,500
Yakima R.	400	3,200	8	Zillah, Wash.	1:125,000

* Width of stream probably less than that responsible for meander belt.

This is indicated by the broken line in Figure 13. The variation of ratio with changes of width may be due either to the fact that the relation between width of meander belt and width of stream is not truly expressed by a linear equation or to a systematic error in measuring the quantities on a topographic map. Neither of these points, however, is thought to be of sufficient magnitude to alter the general relations indicated by the data.

The moderate deviation of the ratios from the mean is a measure of the degree of exactness of the relationship. With two exceptions the ratios of width of meander belt to width of stream fall below 30:1. The exceptions (the Red and Wichita rivers) have ratios of 56 and 39 respectively, but examination of the topographic maps strongly suggests that the meander belt in these cases was not developed by the low water stage the width of which is given on the map.

INCISED MEANDERS

Definition of incised meanders.—By incised meanders is here meant those meanders whose free movement is restricted by the narrowness or absence of a floodplain. Such a definition includes the one-cycle incised meanders described by Winslow (1893, p. 31-32) as developing from the irregular bends of a stream simultaneously with its downcutting. In terms of the fluvial cycle the development of a floodplain is limited to the periods during which lateral cutting predominates over downward cutting—namely, the mature and old age stages of the cycle. The stage at which fully developed meanders may first be formed—when the floodplain is opened out to a width equal to the limiting width of the meander belt—is designated as the beginning of full maturity and marks the transition from early to full maturity (Johnson, 1932, p. 486). In the case of an

incised meandering stream all evidence that meanders were incised, or developed during incision, is removed by full maturity. Incised meandering streams are, therefore, those meandering streams which are in youth or early maturity.

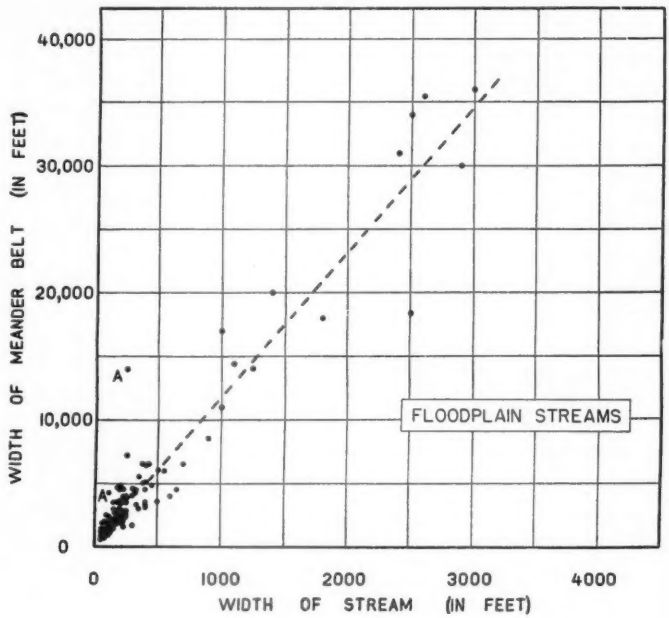


FIGURE 13.—*Meander ratios of streams on floodplains*
A, A represent abnormally large ratios of the Red and Wichita rivers.

Mechanics of meander development.—The growth of meanders on a floodplain is retarded by a decrease in centrifugal force and stopped completely by a cut-off resulting from intercision or overflow across the meander neck in time of flood. The significance of gradient in the two types of meanders can best be appreciated by recalling their relations to the cycle of stream development. Since incised meandering streams are in an earlier stage than those on floodplains, they would as a rule have higher gradients and, other things being equal, higher velocities.

The question may be raised as to how the increased velocity, together with the proportionately decreased cross-sectional area necessary to maintain a constant discharge, will effect the size of meanders developed by incised streams. From equation (12) it can be seen that the centrifugal

force varies as the first power of the cross-sectional area but as the square of the velocity. Therefore the change in velocity is of greater importance in determining the centrifugal force exerted on the meander than is the

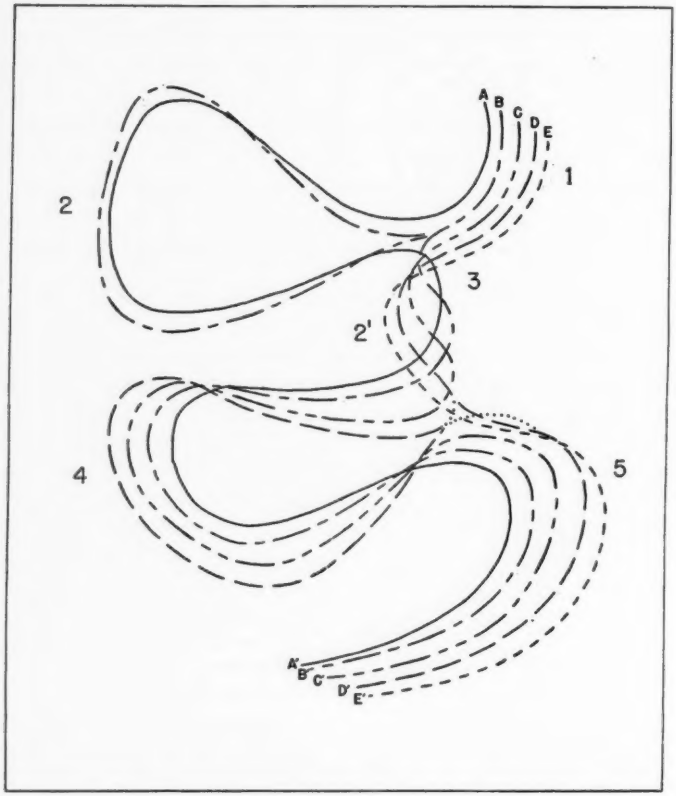


FIGURE 14.—*Elimination of meanders*

change in the cross-sectional area. The increased velocity accompanying incision will, thus, result in increased centrifugal force which will decrease with meander growth at the same rate as in floodplain streams. If the meanders are spaced sufficiently far apart along the stream to permit their full development, incised meanders will actually grow to a size greater than those developed by a stream of the same discharge flowing on a floodplain.

It may be asked to what extent meanders will be respaced as a result of increased centrifugal force, to permit their maximum development. The deduced effects of such a change are presented in Figure 14. The initial stage (AA′, Fig. 14) of meanders 1, 2, 3, 4, and 5 is indicated by the solid line. With sufficiently large increase in either the velocity or the discharge, the resulting increase in centrifugal force will be effective on each meander of the stream. Since the centrifugal force varies inversely as the radius of curvature it will be most effective on those curves which have a small radius of curvature at the time of the increase or those that acquire a small radius of curvature soon afterward as the result of a cut-off. With increased centrifugal force meander 2 will soon be cut off, and growth of the sharp bend thus formed will be accelerated more than in the case of the other curves having a larger radius of curvature (BB′, Fig. 14). The new curve, designated as 2′, will also tend to have a longer radius of curvature than if the increase in centrifugal force had not taken place. The migration of this newly formed curve sharpens meander 3, thus accelerating its growth and shortening the length of time before meander 4 will be cut off (CC′, Fig. 14). If the increase in centrifugal force has been sufficiently large the stream will not be able to develop a small curve to replace meander 4 but will continue directly from meander 2′ to meander 5 (DD′, Fig. 14). This will result in the elimination of two of the five meanders originally present in this stretch of the river. Such an adjustment, repeated more than once if necessary, will after a period of time establish a more stable system of meanders. They will be spaced in such a way that each cut-off (except where two cut-offs take place simultaneously or when migration is retarded by irregularities of rock structure) will result in the formation of a new meander which will take the place of, and as a rule grow to the size of, the one replaced. If the increase in centrifugal force is not sufficiently great or the meanders not spaced sufficiently close together to bring about this type of adjustment, it seems likely that the stream will by less spectacular methods, probably by more intensive erosion on some meanders than others, bring about a respacing of the meanders until they are more nearly in adjustment with the new conditions. As a result of reduction of the number of meanders in a given distance along the river axis, each of the meanders will be able to grow to a greater size before being cut off.

Some cut-offs limiting the growth of meanders on a floodplain take place during floods due to overflow across the neck of the meander. The rest are the result of actual intercision of the stream on the two sides of the meander. In the case of incised meanders, overflow across the neck in time of flood is prevented by the bedrock spur projecting

from the valley wall out into the inside of the meander.[6] Thus, with the exception of subterranean diversion through the neck, which may take place under certain specialized conditions, cut-offs can occur only by actual intercision of the stream on the two sides of the meander neck. The additional lateral corrasion necessary to bring about intercision will at the same time result in further widening of the meander belt. The greater centrifugal force exerted on the curves of an incised meandering stream might conceivably bring about intercision in less time than is necessary for the occurrence of a short-cut on a floodplain. But such acceleration of lateral corrasion would normally be more than offset by the fact that bedrock is usually much more resistant than floodplain deposits, and by the further fact that the high banks of an incised meandering stream contribute far more debris for removal than do the low banks of a stream meandering on a floodplain.

Before larger meander ratios are postulated for incised streams on the basis of these conclusions, the behavior of the other member of the ratio—stream width—in the two types of meandering streams should be studied. The width of a stream is a function of its discharge, its velocity, and the configuration of its channel. With a constant discharge a narrowing of the stream may be accomplished either by increasing its velocity so that the same amount of water can pass through a smaller cross-sectional area in a unit of time, or with a constant velocity by deepening the channel. In general, the velocity of incised streams is greater than in streams on floodplain; this tends to decrease the width of the stream, and thereby automatically increase the ratio of width of meander belt to width of stream.

Data on the characteristic channel forms of incised and floodplain streams are not available, but certain deductive statements can be made. The shape of the channel which has the least resistance to turbulent flow is that which has the largest hydraulic radius—namely, one with a semicircular cross-section (Pickels, 1925, p. 107). Streams will tend to form this type of channel. Stream channels may, however, with constant velocity and discharge depart in form from the optimum condition by becoming either deeper and narrower or wider and shallower. Probably the most important factor determining the direction of the departure is the structure and erosional resistance of the material into which the stream is cutting. The writer, however, has not found adequate data on the relation of channel form to character of bedrock, and investigation of this relationship was beyond the scope of this study.

[6] Suggested by Ralph L. Miller. Personal Communication.

Hence it appears that larger meander ratios for incised streams are to be expected for at least three reasons: (1) short-cuts being impossible, meanders will have grown to greater amplitude before growth is stopped by a cut-off by intercision; (2) because of its greater veloc-

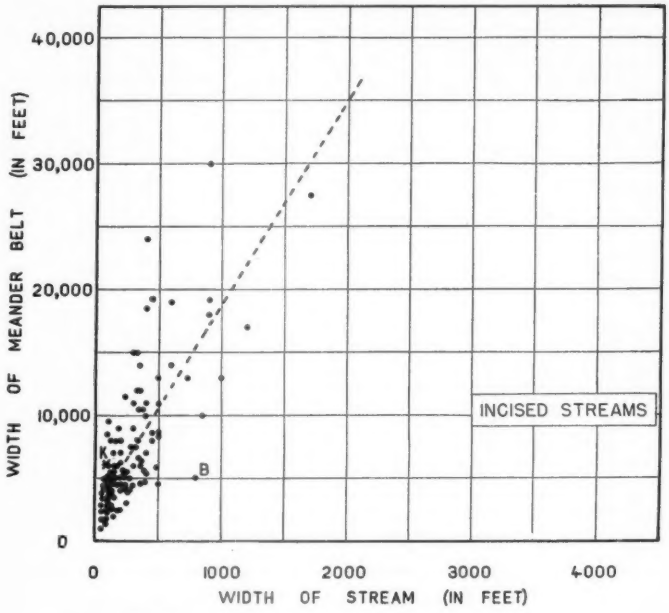


FIGURE 15.—*Meander ratios of incised streams*

B indicates abnormally small ratio of stream whose width is exaggerated on map (Holston River). K is ratio of hypothetical Kickapoo, using width of present stream and the meander belt formed by large meanders of valley.

ity an incised stream will be narrower than a stream of the same discharge flowing on a floodplain, thus automatically increasing the ratio; and (3) the centrifugal force exerted against the concave bank of a meander will be increased as a stream of given discharge attains a higher velocity, thus tending toward more widely spaced meanders which can grow to a greater size before being cut off. In view of these conclusions data giving actual relationships of width of meander belt to width of incised streams will be examined.

Relation of meander size to stream size.—On the basis of 30 measurements Jefferson (1902, p. 382) has given 30:1 as the mean ratio of the

TABLE 3.—*Meander Ratios of Incised Streams*

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
Allegheny R.	500	8,500	17	Franklin, Hilliards, and Foxburg. Penn.	1:62,500
Antietam Cr.	90	1,800	20	Hagerstown, Md., Pa.	1:62,500
Arkansas R.	900	18,000	20	Skedee, Okla.	1:62,500
Ashtabula R.	50	2,800	56	Conneaut, O., Pa.	1:62,500
Barren R.	160	8,000	50	Scottsville and Bowling Green, Ky.	1:62,500
Barren R.	75	5,000	67	Red Boiling Springs, Tenn., Ky.	1:62,500
Belle Fourche R.	120	3,600	30	Aladdin, Wyo., S. D., Mont.	1:125,000
Big R.	150	5,500	37	Bonneterre, Mo.	1:62,500
Black Warrior R. Locust Fork.	150	5,000	33	Birmingham, Ala.	1:125,000
Brazos R.	350	14,000	40	Godley, Texas.	1:62,500
Brazos R.	230	11,500	50	Pickwick, Texas.	1:62,500
Double Mtn. Fr.*.	200	5,000	25	Camp Springs and Rotan, Texas.	1:62,500
Cacapon R.	140	7,000	50	Capon Bridge, Va., W. Va.	1:62,500
Cahaba R.	150	2,000	13	Bessemer Iron District, and Montevallo, Ala.	1:62,500
Canadian R.*.	600	19,000	32	Stonewall, Okla.	1:125,000
Captina Cr.	75	4,500	60	Clarington, O., W. Va.	1:62,500
Cheyenne R.	300	6,000	20	Oelrichs, S. D., Neb.	1:125,000
Cimarron R.*.	330	12,000	36	Yale and Ripley, Okla.	1:62,500
Clarion R.	275	4,000	14	Foxburg, Pa.	1:62,500
Clinch R.	130	3,700	28	Carterton, Va.	1:62,500
Collins R.	360	4,600	12	McMinnville, Tenn.	1:125,000
Colorado R.	400	10,000	25	Spicewood and Mt. Bonnell, Tex.	1:62,500
Colorado R.	150	6,000	40	Bend and Goldthwaite, Texas.	1:62,500
Conestoga Cr.	175	4,500	26	Lancaster, Pa.	1:62,500
Conococheague Cr.	180	9,000	50	Williamsport, W. Va., Md., Pa.	1:62,500
Conodoguinet Cr.	300	7,500	25	Harrisburg and New Cumberland, Pa.	1:62,500
Coosa R.	300	11,000	37	Fort Payne and Anniston, Ala.	1:125,000
Cumberland R.	400	24,000	60	Hermitage and Nashville, Tenn.	1:62,500
Cumberland R.	325	15,000	46	Gainsboro and Carthage, Tenn.	1:62,500
Cumberland R.	400	18,500	46	Monticello, Ky.	1:62,500
Cumberland R.	350	10,500	30	Burnside, Ky.	1:62,500
Caney Fork.	200	6,200	31	Gordonsville, Tenn.	1:62,500
Caney Fork.	450	8,600	19	McMinnville, Tenn.	1:125,000
Delaware R.	350	6,500	18	Milford and Port Jervis, Pa., N. Y., N. J.	1:62,500
Delaware R.	400	5,500	14	Long Eddy, Pa., N. Y.	1:62,500
East Branch.	450	8,000	18	Long Eddy and Walton, Pa., N. Y.	1:62,500
Des Moines R.	400	4,700	12	Boone, Iowa.	1:62,500
Dolores R.	120	2,600	22	Paradox Valley, Colo.	1:125,000
Elk R.	60	3,800	63	Decherd, Tenn.	1:62,500
Elk R.	230	5,500	24	Clay, W. Va.	1:62,500
Elkhorn Cr.	100	9,500	95	Lockport, Ky.	1:62,500

* Width of stream less than that responsible for meander belt.

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
North Fork	125	3,500	28	Frankfort, Ky.	1:62,500
Etowah R.	75	4,000	53	Acworth and Tate, Ga.	1:62,500
Fish Creek	90	4,000	44	Cameron, Ohio	1:62,500
Gauley R.	230	4,500	19	Fayetteville, W. Va.	1:62,500
Grand R.	150	3,500	23	Ashtabula, Ohio	1:62,500
Green R.	275	5,000	18	Mammoth Cave, Ky.	1:62,500
Greenbrier R.	230	5,500	24	White Sulphur Springs, W. Va., Va.	1:62,500
Guadalupe R.	135	5,000	37	Smithson Valley, Tex.	1:62,500
Guyandot R.	130	3,800	29	Gilbert, W. Va., Va., Ky.	1:62,500
Harpeth R.	90	8,500	94	White Bluff, Tenn.	1:62,500
Holston R.	800	5,000	6	Morristown, Tenn.	1:125,000
Holston R.	850	10,000	12	Greenville and Estillville, Tenn.	1:125,000
Honeoye Cr.	80	1,350	17	Rush, N. Y.	1:24,000
Hurricane Cr.	50	2,400	48	Cottondale, Ala.	1:62,500
James R.	110	5,200	47	Shell Knob, Mo.	1:62,500
James R.	330	8,000	24	Buchanan and Eagle Rock, Va.	1:62,500
John Day River	200	4,000	20	Condon, Oregon	1:125,000
Juniata R. Raystown Br.	225	5,100	23	Huntingdon and Mt. Union, Pa.	1:62,500
Kentucky R.	350	12,000	34	Lockport, Ky.	1:62,500
Kentucky R.	375	10,500	28	Frankfort, Ky.	1:62,500
North Fork	110	3,500	32	Buckhorn, Ky.	1:62,500
Kickapoo R.† (restored)	80	6,000	75	Wauzeka, Wis.	1:62,500
Kings R.	125	5,500	44	Shell Knob, Mo.	1:62,500
Levisa Fk.	100	3,500	35	Regina, Va., Ky.	1:62,500
Licking R. South Fork	120	8,000	67	Cynthiana, Ky.	1:62,500
Little R.	90	2,500	28	Appling, Ga.	1:62,500
Little Kanawha R.	150	4,500	30	Arnoldsburg, W. Va.	1:62,500
Little Miami R. East Fork	100	3,200	32	Batavia, Ohio	1:62,500
Mahoning Cr.	130	4,000	31	Rural Valley, Pa.	1:62,500
Maumee R.	230	5,800	25	Paulding, Ohio	1:62,500
Minnesota R.	50	1,000	20	Peever, S. D., Minn.	1:62,500
Monongahela R. (locked)	730	13,000	18	Brownsville and Masontown, Pa.	1:62,500
Monongahela R. (locked)	400	5,500	14	Blacksville, W. Va., Pa.	1:62,500
Muskingum R.	400	11,000	27	Marietta, Ohio	1:62,500
Neosho R.	500	13,000	26	Wyandotte, Okla., Kan., Mo.	1:125,000
New R.	500	4,500	9	Independence and Galax, Va., N. C.	1:62,500
South Fork	250	4,000	16	Wilkesboro, N. C.	1:125,000
Nolin R.	125	2,800	22	Cub Run, Ky.	1:62,500

† Not included in computations of mean.

Stream	A Width of Stream	B Width of Meander Belt	Ratio of B:A	U. S. G. S. Map	Scale
	Feet	*Feet*			
North R.	60	3,300	55	Samantha, Ala.	1:62,500
North Fork R.	300	9,000	30	Mountain Home, Ark., Mo.	1:62,500
Obey R.	140	6,000	43	Lillydale, Tenn.	1:62,500
Ohio R.	1,700	27,500	16	Cannelton, Ky., Ind.	1:62,500
Ohio R.	1,200	17,000	14	Ravenswood, W. Va., O.	1:62,500
Ohio R.	900	19,200	21	Clarington, W. Va., O.	1:62,500
Osage R.	900	30,000	33	Versailles and Tuscumbia, Mo.	1:125,000
Potomac R.	500	11,000	22	Williamsport and Martinsburg, W. Va., Md.	1:62,500
Potomac R.	300	15,000	50	Pawpaw, Md., W. Va., Pa.	1:62,500
North Fork	110	4,300	39	Elk Garden, W. Va., Md.	1:62,500
South Fork	200	8,000	40	Hanging Rock and Flintstone, W. Va.	1:62,500
Pound R.	90	2,300	25	Clintwood, Va., Ky.	1:62,500
Powell R.	100	4,700	47	Middlesboro, Tenn., Ky.	1:62,500
Red R.	250	3,000	12	Fargo, N. D., Minn.	1:125,000
Rockcastle R.	300	4,400	15	London, Ky.	1:125,000
Rum R.	175	2,400	14	St. Francis, Minn.	1:62,500
Salt R.	150	5,000	33	Taylorsville, Ky.	1:62,500
Schuylkill R.	360	6,000	17	Phoenixville, Pa.	1:62,500
Shenandoah R.	350	6,500	18	Berryville, W. Va., Va.	1:62,500
North Fork	400	7,000	17	Luray, Va.	1:125,000
South Fork	500	8,600	17	Luray, Va.	1:125,000
Shiawassee R.	60	1,700	28	Chesaning, Mich.	1:62,500
Shoal Cr.	200	4,500	22	Muscle Shoals, Ala.	1:62,500
Stone R.	100	6,000	60	Hermitage, Tenn.	1:62,500
Susquehanna R. West Branch	250	4,000	16	Clearfield and Karthaus, Pa.	1:62,500
Tennessee R.	1,000	13,000	13	Loudon, Tenn.	1:125,000
Twelvepole Creek	100	3,500	35	Guyandot, W. Va., O.	1:62,500
Valley Cr.	100	2,800	28	Yolande, Ala.	1:62,500
Verde R.	200	2,400	12	Turrett Peak, Ariz.	1:125,000
White R.	600	14,000	23	Yellville, Ark.	1:125,000
White R.	220	5,300	24	Forsyth, Mo.	1:125,000
White R.	200	7,000	35	Shell Knob, Mo.	1:62,500
Wolf R.	100	4,000	40	Byrdstown and Lillydale, Tenn.	1:62,500
Yampa R.	125	2,500	20	Axial and Monument Butte, Colo.	1:62,500
Youghiogheny R.	300	7,500	25	Pittsburgh, Pa.	1:62,500

width of meander belt to width of incised streams. Data assembled by the writer in precisely the same way as those for the floodplain streams reveal a similar relation. Measurements given in Table 3 and in Figure 15 strongly suggest a straight line relation. The mean

straight line ratio is found by least squares to be 41:1 for streams 100 feet wide, 18:1 for streams 1000 feet wide, and 17.5:1 for streams 2000 feet wide. The characteristics of the streams shown on some of the older topographic maps of the United States strongly suggest that the actual stream width is only a fraction of the width shown on the maps. The effect is to reduce the size of the meander ratios. The greater number of these early maps are in regions of incised streams—namely, plateaus and mountains having strong relief. The reduction of ratios due to this source of error notably decreases the difference between the ratios of the incised and the floodplain meanders. As would be expected from the greater variety of conditions involved, the deviation of ratios from the mean is much greater in the case of incised streams than for streams on a floodplain. The ratios of incised streams vary from 6:1 to 95:1.

MISFIT FEATURES OF THE KICKAPOO RIVER

Identification of misfit meander scarps.—Before discussing the cause of the misfit character of the Kickapoo it is necessary to establish the fact that the meanders of the valley and the meander scarps cut in the valley walls were developed by meanders of a former Kickapoo River. An alternative is that they were developed by irregular bends which had no particular significance. In distinguishing between bends and meanders of a river Davis (1924, p. 56) has described bends as original curves that result from irregularities in the surface and which have no relation to the size of the stream. He defined meanders as regular curves whose radius of curvature is adjusted to the size of the stream. It is important in this connection to consider only those curves in which sufficient arc is preserved to determine accurately the radius. As has been pointed out the radius of curvature varies in length throughout meander development. Since the river meanders responsible for some of the scarps may not have been fully developed at the time the river became misfit, particular significance would attach to high level meander scarps which were fully developed and abandoned before the stream acquired its misfit character. Thus the constancy of radius, and the smoothness of the meanders of the valley, and the crescentic scarps cut in the valley sides (particularly those abandoned above present stream level) are apparently the most satisfactory criteria for determining which have been formed by meanders of the former stream and which have been cut by the irregular bends of such a stream.

The crescentic scarps of the Kickapoo Valley, including several high level meander scarps, are extremely smooth and of remarkably uniform curva-

ture. A similar smoothness and uniformity of curvature is exhibited by many meanders of the Kickapoo Valley. They are, therefore, interpreted as true meander forms developed by an earlier Kickapoo River.

Cause of misfit character.—If the meander scarps and valley meanders are to be considered true meander forms, the change in the meandering habit of the stream must be explained. As set forth by Davis (1896) the concept of misfit streams involved the development of meander forms by a stream which was subsequently deprived of part of its flow, due to a diversion of former headwaters or to climatic change. Davis (1913, p. 4) later called attention to the suggestion of Dr. Otto Lehmann of Vienna that a mature stream may acquire a misfit character by loss of volume as a result of underflow through alluvium on the valley floor.

Diversion of former headwaters of the Kickapoo River since the dissection of the Dodgeville peneplane can be eliminated as a possible cause of the misfit character by an examination of the Kickapoo basin. The divide surrounding the headwater portions of the Kickapoo basin is formed by the highest areas along the crest of the Prairie du Chien or "Magnesian" cuesta and rises 350 feet above the valley just a few miles to the south. It is extremely regular, and the absence of any sag whereby extensive headwaters might have formerly joined the Kickapoo indicates that no diversion of flow could have taken place later than the uplift and dissection of the Dodgeville peneplane.

It is possible that the Kickapoo may have had more extensive headwaters that were diverted prior to the dissection of the Dodgeville peneplane by the development of the subsequent lowland on the weak St. Croixan sandstones just north of the Kickapoo basin. Such a diversion may have materially reduced the discharge of the Kickapoo and, assuming the river was flowing in meanders appropriate to its size on the peneplane, may have resulted in the development of a misfit stream. It is further possible that with uplift of the Dodgeville peneplane and incision of the small misfit meanders, down-valley migration of these meanders (which were themselves following the course of the larger meanders that existed prior to the diversion) may have produced the meandering valley now occupied by the Kickapoo, as well as the several crescentic scarps cut in the valley walls. The maintenance of the original meander pattern by the small meanders with sufficient regularity throughout a few hundred feet of incision to produce the extremely uniform features of the Kickapoo Valley appears to the writer unlikely.

Another type of diversion frequently found in or near glaciated areas is the temporary diversion of glacial meltwater into a drainage area. As previously mentioned the drainage basin of the Kickapoo River is entirely

within the Driftless Area and whatever glacial water found its way into the Kickapoo must have come over the high divide mentioned above, or escaped through the sandstone of the divide. Flow of glacial meltwaters over the divide is negated by the evidences already discussed, and no evidence indicating flow through the sandstone has been discovered.

The loss of discharge due to flow through the alluvium, according to what Davis (1913, p. 4) has called Lehmann's Principle, is important in some streams in semi-arid regions and is especially applicable to streams whose valley floors consist of boulder beds covered with finer material. In the Kickapoo the alluvial filling is 50 to 125 feet deep and consists, insofar as can be determined from well records, of fine sand and silt with increasing quantities of gravel toward the bottom of the fill. Although no further data are available on this point, underflow must be considered at least a partial explanation for the misfit features of the Kickapoo.

The possibility of climatic change is of regional rather than local consequence. If we accept Trowbridge's (1935, p. 62, 75) suggestion that practically all of the present dissection of the Driftless Area is Pleistocene (post-Nebraskan) in age, the cutting of the present valleys would have been greatly affected by the climatic conditions during that time. There is, however, some difference of opinion as to the climatic conditions of the Pleistocene in regions near the ice margin. The lower temperature may have reduced the amount of vegetation protecting the land surface, thus increasing the amount of precipitation that is disposed of as run-off. A similar effect would be brought about by the decreased evaporation at the lower temperatures. Opinions differ, however, as to the amount of precipitation at that time. Martin (1932, p. 133) has expressed the opinion that

> "the precipitation over the whole Driftless Area was undoubtedly increased by the presence of the surrounding continental glacier and the cold that it brought."

On the other hand, if the prevailing winds were anti-cyclonic blowing from the ice mass to warmer areas, as has been suggested by Hobbs (1926) and favored by many European glaciologists (Bryan, 1928), the precipitation would be decreased due to the increasing capacity of the atmosphere to hold moisture. Such climatic changes would be effective only during the glacial epochs of the Pleistocene; conditions of the interglacial epochs would presumably be more nearly like those of the present day. Because of the regional character of climatic change, the appearance of other streams in the region is of particular significance. Although other streams do not exhibit the prominent high level crescentic scarps that constitute the best evidence of a misfit stream, some of the valleys do possess meander-like curves that suggest a larger meander pattern

than that of the present streams. In view of these facts the possibility of climatic change as an explanation of the misfit character of the Kickapoo cannot be discarded.

Thwaites (1935, p. 106) discards all explanations involving "loss of volume",[7] and states that the misfit character of the Kickapoo is due to the fact that the stream is now flowing on a floodplain of silt and sand instead of its former rock bed. He fails to explain why meanders of streams flowing on alluvium should be smaller than those of streams flowing on bedrock. On the basis of data on incised and floodplain meanders given in this paper one may, however, deduce that a stream might acquire a misfit relationship without a diminution of discharge. This would be accomplished by first having meanders and associated meander forms, such as scarps, cut-banks, and slip-off slopes, grow larger during incision than in the same stream on a floodplain. Then, if aggradation is introduced by some external factor before the stream has normally begun to develop its floodplain, the current may be expected to abandon the large meander forms produced during incision and begin new meanders whose size is within the normal limits of growth for streams of that size on a floodplain.

It may be asked, then, whether the normal history of an incised meandering stream will not include the acquisition of a misfit character as soon as it widens out the valley floor and develops a floodplain. After sufficient down-cutting the stream will begin to develop a floodplain of its own, but the acquisition of misfit characteristics by such normal development seems doubtful. It is more probable that as the stream approaches a graded condition, at which time the floodplain begins to develop, the change in meander characteristics will be gradual; older meander forms, except high level meander scarps, will be largely destroyed before the change has been completed. Consequently, at no time will there be a sharp contrast between the forms developed during incision and those developed after the formation of a floodplain.

Misfitness of this type could come about only under conditions which permit the growth of meanders during incision, whatever those conditions may prove to be. In the Kickapoo there is a sharply incised meandering valley and extensive aggradational deposits on the valley floor. This explanation, therefore, seems to offer a possible solution of the misfit character of the Kickapoo. It should be noted in applying it to the Kickapoo that the stream at the time of incision need not have had so

[7] ". . . the change in size of curves [of the Kickapoo] is explicable by the fact that the stream is now on a floodplain of silt and sand instead of the rock bed it had before the glaciation of the surrounding region with consequent raising of the level of all the streams. The Kickapoo is truly an "underfit river" but the cause is not loss of volume by diversion or other means."

small a meander ratio as it does at the present time; on the contrary, the ratio may have been any size within a maximum of at least 30:1, the largest meander ratio observed on floodplain streams.

The appearance of the Kickapoo Valley at the time the large meander

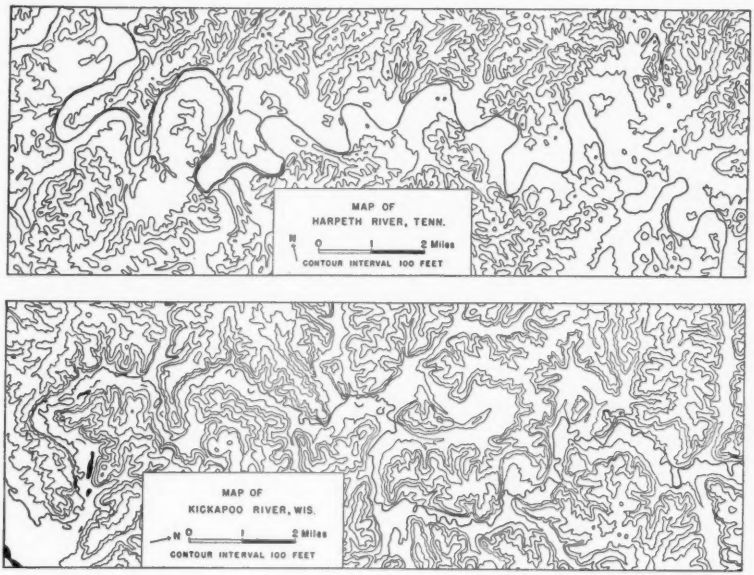


FIGURE 16.—*Comparison of valleys of Harpeth and Kickapoo rivers*

Showing meander forms being developed by Harpeth River and abandoned meander-like curves of Kickapoo valley. Topography from U. S. Geol. Survey. Width of streams not to scale.

forms were being developed can better be appreciated by comparing it with the remarkably similar forms now being developed by the Harpeth River, Tennessee, a stream only slightly larger than the Kickapoo (Fig. 16). The width of the meander belt formed by the large curves of the Kickapoo Valley and the width of the present stream have been used to compute the meander ratio of the Kickapoo when the large curves were developed. The ratio is 75:1, well below the maximum ratios recorded on incised streams—i.e., the 95:1 ratio of the Harpeth River.

It appears, therefore, that the misfit condition of the Kickapoo may best be explained as the result of either underflow through the alluvium, climatic change, aggradation of an incising stream, or of a combination of these. Available data permit the discarding of only one of the explanations—namely, the diversion of former headwaters in the present cycle.

Diversion on the Dodgeville peneplane is regarded by the writer as very doubtful.

Preservation of valley meanders.—Examination of different parts of the Kickapoo Valley reveals considerable variation in appearance of the meander scarps and meander-cores,[8] as well as of the meandering pattern of the valley. The variation is believed to be largely structural. To appreciate the situation it should be recalled that the dip of the strata is in approximately the same direction and only slightly greater than the gradient of the stream. Thus, the base of the Prairie du Chien dolomite, which caps the valley sides throughout the length of the Kickapoo, rises from slightly above stream level at its mouth to an elevation of more than 300 feet above the stream in the upper portion of the valley. As the base of the formation rises toward the north these meander forms are less well protected by the capping of dolomite.

On the basis of preservation of valley meanders, the valley may be divided into four sections. South of Gays Mills the base of the dolomite extends down to within less than 75 feet of the present stream level. It apparently has been sufficiently resistant to retard the destruction of meander-cores and inter-meander spurs by lateral and down-stream migration of the meander until a late stage of incision. As a result they have not been modified appreciably. This section contains the best developed meanders of the entire valley and includes the two cut-off loops, Citron and Haney valleys (Fig. 8 and Pl. 6, fig. 2).

The next section of the valley is somewhat longer than the first and includes that part of the river between Gays Mills and LaFarge (Fig. 8 and Pl. 5, fig. 1). Here the valley has cut to a greater depth in the weak sandstones underlying the Prairie du Chien dolomite, and the meander forms are not so well preserved. A greater length of time has elapsed since the stream cut through the dolomite, so there has been more opportunity for widening of the valley and modification of the meander scarps and inter-meander spurs by slope wash, creep, and migration of the meanders. The absence of a resistant caprock has permitted the almost complete reduction of the meander-core 1¼ miles southwest of Soldiers Grove.

The third part of the valley extends from LaFarge to 3 miles north of Ontario, a distance of 12 miles. The outstanding feature of this section is the development of rock terraces which form the floor of an outer valley 400 feet deep and a mile in width, and below which the stream is incised 100 to 150 feet in large well developed meanders (Fig. 9 and Pl.

[8] The term "meander-core" was proposed by A. M. Davies (1923, p. 88) to designate "the central hill encircled by the meander."

FIGURE 1. TERRACE MATERIAL
Gully near valley side 1¼ miles southwest of Soldiers Grove. Shows terrace material with stratification parallel to sloping surface of terrace.

FIGURE 2. FILLED TRIBUTARY VALLEY
One and a half miles north of LaFarge on west side of valley.

VALLEY FILL

FIGURE 1. KICKAPOO VALLEY NEAR GAYS MILLS
Looking upstream. Lighter areas in foreground and in left distance are alluvial terraces. Dark area is present floodplain. Transverse profile of terrace is visible in center distance.

FIGURE 2. VALLEY FILL
Exposure one mile west-southwest of Soldiers Grove. Extensive gravel layer interbedded with finer stratified sand.

VIEWS IN LOWER KICKAPOO VALLEY

6, fig. 1).[9] Although it follows this large meandering pattern, the present stream has formed a series of small meanders. Exposures along the inner valley of the stream show that the rock terraces are underlain by the Ironton sandstone member of the Franconia formation (Pl. 7, fig. 1). This is a resistant stratum within the generally non-resistant St. Croixan sandstones that underlie the Prairie du Chien dolomite.

The formation of these terraces may be explained as a result of (1) a pause in downcutting, in which a broad valley floor was developed, or (2) the removal of the less resistant material overlying the Ironton sandstone by slope wash and creep on the meander spurs. The pause in downcutting may have been due to a close approach to base level and hence of cyclical significance, or to the presence of a resistant barrier downstream. The Prairie du Chien dolomite now cut through by the Wisconsin River near its mouth has been suggested by Trowbridge (1921, p. 111-112) as the barrier responsible for the terraces. Thwaites (1935, p. 106) attributes the presence of these terraces to the resistant layer of sandstone. The irregular surface of the terrace together with the fact that no comparable terrace forms are found in other parts of the valley leaves the hypothesis of their being the result of a pause in downcutting without convincing support. On the other hand, with a strong contrast in rocks and an inner trench meandering from one side of the valley to the other, conditions seem very favorable for complete removal of material above the resistant sandstone layer by slope wash and creep. The removal of this weaker material would be accelerated by migration of the meanders during incision. That such has been the case in the upper Kickapoo Valley is suggested by the many cut-banks and slip-off slopes and by the abundant stream-worn gravel found on several of the terrace slopes. It is, therefore, concluded that the terraces between LaFarge and Ontario are the result of the removal of material above the resistant Ironton sandstone by slope-wash and creep, aided by the lateral and down-valley migration of the meanders. South of LaFarge the valley is straighter, and the top of the Ironton sandstone more nearly approaches the river level. For this reason the rock terraces are less pronounced and with very few exceptions are entirely masked by the accumulations of talus and slop-wash along the valley sides.

The part of the valley above Ontario comprises the fourth section and is quite different from the others. Absence of the broad outer valley is explained by the fact that near Ontario the Kickapoo is joined by two tributaries, Bear and Morris creeks, each of which is more than half the

[9] Because of generalizations in parts of the topographic map of the LaFarge quadrangle, terraces and the meandering course of the valley are not shown to full advantage.

size of the main stream. Apparently the smaller streams above this point have not been able to develop the wide meandering pattern which made possible the extensive removal of material above the resistant sandstone by slope-wash and creep farther downstream. Consequently, al-

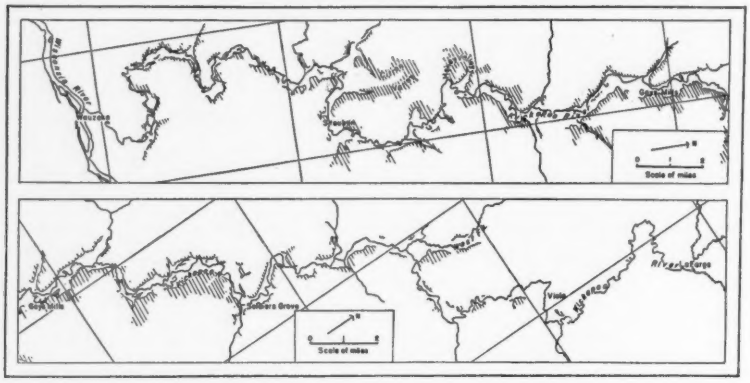


FIGURE 17.—*Map of alluvial terraces in Kickapoo valley*
Terrace surface indicated by diagonal lines. Line along margin of terrace indicates a terrace scarp.

though the Ironton sandstone is present in the valley sides in this section, the terraces are much narrower and more irregular.

AGGRADATION OF THE KICKAPOO VALLEY

General statement.—Throughout the lower 40 miles of the Kickapoo Valley the most conspicuous feature is the accumulation of alluvial material along the valley sides. These accumulations are in the form of pronounced terraces developed by trenching followed by lateral swinging of the river at approximately its present level. Alluvial material also underlies the present river to a depth of 50 to 125 feet, the depth increasing toward the mouth of the stream. As shown in Figure 17, the terraces are fairly continuous along the valley sides and on first sight appear to be remnants of a former floodplain. Closer examination, however, reveals that these features do not have a relatively flat surface sloping generally downstream as is typical of simple river terraces. Instead, the slope of the surface, as seen in profile across the valley, increases from practically zero where they extend well out into the center of the valley to 20 to 25 degrees where they merge smoothly with the valley sides.

The grading of the slope of the terrace surface to the valley sides indicates that the terrace surface was developed by material moving down the valley sides toward the center of the valley, as if by slope-wash or

creep. It remains to be determined, however, whether the entire deposit is developed by slope wash or whether it consists principally of fluvial material covered by a wedge of wash from the valley sides. Further data on this point are obtained by examining the composition of the terraces.

Material in terrace deposits.—The composition of the terrace deposits of the Kickapoo Valley can be seen in small scattered exposures throughout the terraced area. Most of these sections are exposed in gullies that have been cut back from the present floodplain of the stream toward the valley sides. The terraces are composed of sand containing angular fragments of chert. The material is remarkably homogeneous, both in texture and lithology. The homogeneity may be explained by the fact that throughout its length the Kickapoo River has cut through the resistant Prairie du Chien dolomite, which caps the immediate valley walls, and has cut into the underlying St. Croixan series to a depth of less than 75 feet at the mouth of the river and to more than 300 feet in the upper part of the valley. Most of the St. Croixan series, consisting of sandstones and sandy glauconitic shales, is not well indurated, and the disintegrated fragments of these rock types make up a very large percentage of the deposits. The chert fragments are derived from the overlying Prairie du Chien formation. The presence of the same constituents in but slightly varying amounts throughout the drainage basin accounts for the homogeneity of the material. Because of the homogeneity it is impossible to determine from lithology the percentage of locally derived materials and the percentage of the material that has been carried some distance by the stream.

Stratification of the deposits, as shown by alternating light and dark layers, is somewhat irregular but, in general, parallels the surface of the terrace (Pl. 7, fig. 3; Pl. 4, fig. 1). The light layers vary from 1 to 5 inches in thickness, whereas the dark layers are usually less than an inch thick. The darker layers have been cemented with iron oxide so that they are appreciably more resistant than the lighter layers. Consequently, they stand out in miniature ledges between which the loose sand of the lighter layers has been partially removed. In a number of places the stratification of the unconsolidated alluvial material has been badly disrupted apparently as a result of slumping. In other places, mostly near the valley sides, there are occasional gravel lenses less than a foot in thickness which seem to be the result of concentrated wash similar to that taking place on the present valley sides as shown in Plate 6, figure 3.

In three or four exposures in the lower Kickapoo Valley there are either extensive gravel lenses, clay beds, or cut-and-fill stratification indicative

of fluvial conditions (Pl. 5, fig. 2; Pl. 7, fig. 2).[10] With these few exceptions there is no evidence that the alluvial material in the valley has been deposited or reworked by the stream, and in only one exposure is there considerable thickness of clay suggesting ponding of the stream for any length of time. The exposure of clay is located near the power dam along the west side of the valley approximately 1½ miles north of LaFarge (Pl. 4, fig. 2). At this locality 15 feet of clay together with the underlying 25 feet of coarser alluvial material has been deposited during aggradation in the mouth of a small tributary. The tributary has migrated laterally before again cutting downward and thus has cut a new channel to one side and has left the valley filling intact. With these few exceptions the even texture, the somewhat irregular inclined stratification, and the absence of fluvial characteristics seems to indicate that the material has been deposited largely as a result of slope wash and creep from the valley sides. In some exposures the stratified deposits of slope wash are covered by a fine structureless material probably of loessal origin. The loess is in places as much as 5 to 6 feet in thickness.

Extent of aggradation.—If the filling of the Kickapoo Valley was accomplished primarily by wash from the valley sides, it follows that the valley would be filled to greater depths near the sides. The highest exposure above the present floodplain is just southeast of Soldiers Grove where Baker Creek cuts close to the valley side just before joining the Kickapoo. The top of the scarp is 80 feet above the stream and judging from the slope of the terrace surface the fill extends at least 40 feet higher. It is impossible from the remnants that are preserved to determine accurately the height to which the fill reached in the center of the valley. However, by projecting the surface of the remnant into the valley with a slope similar to that found on the more extensive remnants, it is concluded that the filling must have been about 60 feet above the present floodplain. Incidentally, this figure approximates the height above the present stream level of the floor of Citron Valley, the large cut-off of the Kickapoo Valley near Steuben (Fig. 17). These two higher remnants of the valley fill, as well as another on the west side of the valley 4 miles south of Gays Mills, do not conform to the general impression that all of the alluvial terraces were graded to the same level. The fact that no filled channels have been discovered in the area of rock terraces in the upper Kickapoo Valley suggests that aggradation was insufficient to cover the surface of these terraces. This would seem to limit the fill

[10] The exposures are: North side of river, 1 mile N. of Wauzeka; West side of river, 4 miles S. of Gays Mills; South side of river, 1½ miles WSW. of Soldiers Grove.

FIGURE 1. KICKAPOO VALLEY BETWEEN LEFARGE AND ONTARIO
Looking south from Mt. Pisgah showing broad outer valley, floor of which is surface of rock terraces shown as cleared fields. The Kickapoo is incised below this surface in a meandering pattern indicated by timbered areas. Photograph by F. T. Thwaites.

FIGURE 2. HANEY VALLEY
Hill in left middleground is core of abandoned loop.

FIGURE 3. GRAVEL WASH ON VALLEY SIDE
Along east side of valley 1½ miles north of Soldiers Grove.

VIEWS OF KICKAPOO VALLEY

FIGURE 1. ROCK TERRACE
Photograph showing Ironton sandstone underlying terrace where it is undercut by Kickapoo River 3 miles north of LaFarge.

FIGURE 2. VALLEY FILL
Exposed in high terrace remnant on west side of valley 4 miles south of Gays Mills. Cross-bedded sands and 2½ feet of bluish clay are exposed in the section. Weathered bedrock is exposed in bottom of gully.

FIGURE 3. TERRACE MATERIAL
Exposed in road cut 2½ miles north of Gays Mills. Fresh surface in center shows irregularity of stratification. Weathered exposure on either side shows resistant ledges formed by darker layers.

in the upper part of the valley to not more than 60 feet above the present stream level.

The determination of the number of fillings represented in the Kickapoo Valley presents other difficulties. The material making up the ter-

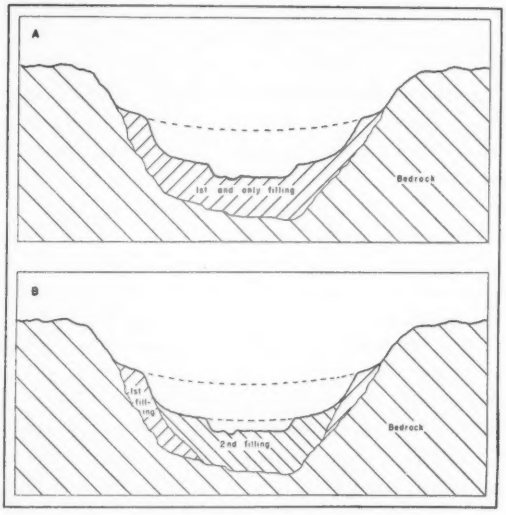


FIGURE 18.—*One and two valley fillings*

races in the valley is apparently the same, as would be expected from the limited types of material available in the basin. Furthermore, all have apparently been formed chiefly by wash from the valley sides. It is, therefore, difficult to tell whether all the terraces have been developed by partial removal of one valley fill as shown in Figure 18A or whether they are the result of two or more periods of filling separated by periods of valley trenching as in Figure 18B. Since the terraces of the Kickapoo Valley are similar in all visible particulars the only method of determining the number of fills represented would be to observe contacts between the two fillings. None has been observed, hence both possibilities must be considered. The question of how many fills are represented in the Kickapoo Valley can probably best be answered by a study of the more distinctive glacial deposits of the Wisconsin River to which the Kickapoo is tributary. MacClintock's (1922, p. 688) study of the material making up these terraces points to repeated filling and excavation of the valley connected with successive stages of Pleistocene glaciation.

Terrace correlation.—The character of the fill and the sloping surface of the terrace remnants raise a question as to what method should be used in their correlation or what principles should govern the correlation of slope-wash terraces. Terrace correlation normally consists of matching together remnants formed at the same stage of a stream's history. In order to ascertain which remnants should be matched it is necessary to determine the level at which the stream was flowing when the terraces were cut. In the case of terraces cut on rock the level is indicated by the beveled surface of the bedrock. A similar horizon may sometimes be apparent in terraces cut on alluvial or unconsolidated material. In any case the correlation must be made between significant elevations on the terrace remnants.

One position on the terrace commonly taken as a significant elevation is the top of the frontal scarp. In slope-wash terraces having smoothly graded sloping surfaces, such as those in the Kickapoo Valley, this elevation is variable and is determined by the extent to which the terraces have been cut back by the stream swinging at a lower level. It thus cannot mark the level of the stream to which the wash slopes were graded. The manner in which such accumulations of slope wash affect the height of a terrace scarp has been brought out very clearly in Mackin's (1937, p. 828-833) study of Shoshone River terraces. In these terraces, however, the layer of channel gravel underlying the slope wash and overlying the beveled bedrock surface marks the stream level at the time the terrace was cut.

Unfortunately no such gravel or beveled surface is to be found in the Kickapoo terraces. Instead, with the few exceptions noted, the material exposed above the present stream level is apparently due almost entirely to wash from the valley sides. It may be possible, however, to get some idea as to the level to which the wash slopes were graded by examining the range of elevation and the slope of the terrace surfaces. The elevations of the frontal edge of the terraces are shown in Figure 19.

As stated earlier most of the terraces give a definite impression of being remnants of a continuous accumulation along the valley side. Making allowance for variation in the height of the top of the terrace scarp discussed above, the elevations shown in Figure 19 are not unfavorable to this interpretation. In some places the material seems to be somewhat concentrated as small alluvial fans at the mouths of the side ravines but, in general, it has apparently been distributed quite evenly along the valley sides.

The extremely gentle slope of the remnants near the center of the valley seems to indicate that with the few exceptions noted the level to which the slopes are graded is very little below the lowest elevations consis-

tently shown in Figure 19. This is approximately 20 to 25 feet above the present floodplain. The few remnants mentioned as representing a possible limit of aggradation were graded to a level approximately 60 feet above the present floodplain.

According to MacClintock (1922, p. 681) the higher Wisconsin ter-

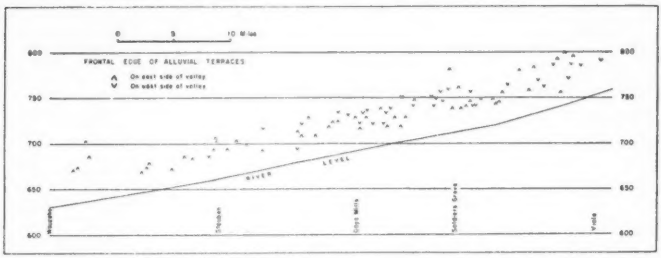


FIGURE 19.—*Elevation (in feet) of frontal edge of alluvial terraces in Kickapoo valley*

races of the Wisconsin River valley are 30 to 40 feet above the present floodplain, and the lower Wisconsin terraces are 12 to 15 feet above the present floodplain. The presence of two prominent terrace levels in the Wisconsin Valley and the indication that wash slopes of the Kickapoo Valley have been graded to two different levels at approximately the same interval above the present floodplain suggest a correlation between the two. Data are insufficient, however, to justify more than a suggestion. Unless the rock terraces found between LaFarge and Ontario be included, there are apparently no terraces in the Kickapoo Valley that may be correlated with the higher pre-Wisconsin terraces of the Wisconsin Valley described by MacClintock (1922, p. 673) and Thwaites (1928, p. 629-631).

One significant point brought out in Figure 19 is the presence of only one terrace at each point along a valley side. In view of the prominence of the one terrace level special effort was made to locate less conspicuous lower terraces, but in only two or three scattered places was any such lower terrace found. This seems to indicate that since trenching itself in the alluvial deposits the stream has remained at approximately its present level for sufficient time to accomplish as much lateral migration as at any time during the trenching.

The present behavior of the stream contrasts sharply with the degrading of the Kickapoo. The swampy condition of the floodplain, the susceptibility of the valley to floods, and the several feet of modern

alluvium revealed by boring on the present floodplain seem to indicate that the Kickapoo, in common with other streams in the Driftless Area, is now aggrading its channel.

Citron and Haney valleys.—Aggradation has been given by Thwaites (1928, p. 627) as an explanation of the two most prominent misfit features of the valley, Citron and Haney valleys (Fig. 8). He suggests that the stream adopted its present course after the valley had been aggraded to the level of a col between the two parts of the valley. A second possibility is that the cut-off occurred during incision of the stream by intercision of two meanders. Although no evidence has been found to eliminate definitely either of the hypotheses, the presence of clay to a depth of at least 13 feet in Citron Valley seems to indicate that the cut off took place before the stream aggraded to within 13 feet of the top of the present fill in the center of the loop.

Cause of aggradation.—In view of the extensive glacial outwash deposits in the Wisconsin Valley, it is natural to relate the nonglacial deposits of the Kickapoo Valley to glacial conditions. The most obvious explanation for the presence of the deposits in the Kickapoo Valley is that aggradation of the Mississippi and Wisconsin rivers forced the Kickapoo to aggrade. The nature of the aggradation that would be expected in the Kickapoo Valley depends to a large extent on the rate of aggradation in the Wisconsin Valley and the supply of material available in the Kickapoo Valley. If aggradation of the Wisconsin Valley was sufficiently rapid, the Kickapoo would not have been able to keep pace, and as a result a lake would have been formed in the Kickapoo Valley. If, on the other hand, the aggradation of the Wisconsin Valley was slower, the Kickapoo would have been able to maintain a stream course during the aggradation. The absence of extensive clay deposits that would result from ponding clearly indicates that the Kickapoo was able to keep pace with the aggradation of the Wisconsin River.

In preceding paragraphs it was concluded that these deposits were largely slope wash which had been reworked in a few places by the stream. This suggests that most of the material making up the fill was locally derived, and that very little was brought down from the upper portions of the valley by the stream. Furthermore, the debris must have accumulated with sufficient rapidity that lateral migration of the stream was largely eliminated.

SUMMARY

The Kickapoo region of southwestern Wisconsin occupies the central part of the Driftless Area and consists of a youthfully dissected plateau developed on gently dipping strata of varying resistance.

Since deposition of Paleozoic and Cretaceous formations in or near the Driftless Area, the erosional history of the area includes the development of one peneplane surface, the remnants of which are found on the highest parts of the region. These remnants bevel the strata on the crests of the cuestas and along the principal divides. From data assembled in the Driftless Area, the lower parts of the upland are interpreted as stripped structural surfaces developed on the backslopes of cuestas. It may be necessary later, however, as a result of work in adjoining glaciated areas, to interpret them as remnants of a second peneplane.

The Kickapoo River is now incised several hundred feet below the upland. During incision of the stream the meanders were enlarged until they exceeded the size found in streams of equal discharge flowing on a floodplain. Thus were the large crescentic scarps cut in the valley walls and large meanders of the valley developed. In contrast to these large meander forms the stream now flows on a floodplain of alluvial material in much smaller meanders, thereby constituting a misfit stream. It is shown that the misfit appearance may result from the sudden introduction of a floodplain by aggradation. Climatic change and underflow through the alluvium are alternative explanations which cannot be discarded.

During incision of the stream rock terraces were developed on a resistant stratum in the upper part of the Kickapoo Valley. The terraces are the result of removal of the overlying nonresistant material from the meander spurs by slope wash and creep, aided by migration of the meanders.

Since incision of the stream the valley has been aggraded in places at least 185 feet above the rock floor of the valley and 60 feet above the present floodplain. The aggradation is attributed to, and took place *pari passu* with, the Pleistocene aggradation of the Wisconsin Valley. The alluvial material has now been partially removed leaving extensive terraces along the valley sides. Examination of surface forms and of material comprising the terraces indicates that the latter has been deposited principally by wash from the valley sides with only limited reworking by the stream.

WORKS TO WHICH REFERENCE IS MADE

Bryan, Kirk (1928) *Glacial climate in non-glaciated regions* (A review of Paul Kessler: *Das eiszeitliche Klima und seine geologishen Wirkungen in nicht vereisten Gebiet*) Amer. Jour. Sci., 5th ser., vol. 16, p. 162-164.

Davies, A. M. (1923) *The abandonment of entrenched meanders: Wye, Evenlode, Cherrvell, Thames,* Geol. Assoc., Pr., vol. 34, p. 81-96.

Davis, W. M. (1896) *The Seine, the Meuse, and the Moselle,* Nat. Geog. Soc., vol. 7. p. 189-202, 228-238. Also *Geographical essays* (1909), Ginn and Co., New York, p. 587-616.

Davis, W. M. (1913) *Meandering valleys and underfit rivers,* Assoc. Am. Geog., Ann., vol. 3, p. 3-28.

——— (1912) *Die erklärende Beschreibung der Landformen.* Leipzig, 2nd Edition 1924.

Exner, F. M. (1931) *Zur Dynamik der Bewegungsformen auf der Erdoberfläche,* Gerland's Beiträge zur Geophysik, Supplementband 1, p. 373-445.

Fenneman, N. M. (1936) *Cyclical and non-cyclical aspects of erosion,* Geol. Soc. Am., Bull., vol. 47, p. 173-186; also Science, n.s., vol. 83, p. 87-94.

Ferrel, Wm. (1859) *The motions of fluids and solids on the earth's surface.* The Mathematical Monthly, vol. 1. p. 140-148, 210-216, 300-307, 366-373, 397-406; vol. 2, p. 89-97, 339-346, 374-390.

Fisher, E. F. (1906) *Terraces of the West river, Brattleboro, Vermont,* Boston Soc. Nat. Hist., Pr., vol. 33, p. 9-42.

Grant, U. S., and Burchard, E. F. (1907) *Description of the Lancaster and Mineral Point quadrangles.* U. S. Geol. Survey, Geol. Atlas, folio 145, 14 pages, 4 maps.

Hershey, O. H. (1896) *Preglacial erosion cycles in northwestern Illinois,* Am. Geol., vol. 18, p. 72-100.

Hobbs, W. H. (1926) *Glacial anticyclones: the poles of atmospheric circulation.* New York. The Macmillan Co., 198 pages.

Jefferson, Mark (1902) *Limiting width of meander belts,* Nat. Geog. Mag., vol. 13, p. 373-384.

Johnson, Douglas (1932) *Streams and their significance,* Jour. Geol., vol. 40, p. 481-497.

Leighly, J. B. (1932) *Toward a theory of the morphologic significance of turbulence in the flow of water in streams,* Univ. of Calif., Pub. in Geog. vol. 6, p. 1-22.

——— (1934) *Turbulence and the transportation of rock debris by streams,* Geog. Rev., vol. 24, p. 453-464; (Abstract) Am. Geophys. Union, Tr. 1934, p. 453-454.

MacClintock, Paul (1922) *The Pleistocene history of the Wisconsin river,* Jour. Geol., vol. 30, p. 673-689.

Mackin, J. Hoover (1937) *Erosional History of the Big Horn Basin, Wyoming,* Geol. Soc. Am., Bull., vol. 48, p. 813-894.

Martin, Lawrence (1911) *Physical geography of the Lake Superior region,* U. S. Geol. Survey, Mon. 52, p. 85-117.

——— (1932) *Physical geography of Wisconsin,* Wis. Geol. and Nat. Hist. Surv., Bull. 36, 2nd Edition.

Pickels, G. W. (1925) *Drainage and flood control engineering.* McGraw-Hill, New York.

Rudski, M. P. (1911) *Physik der Erde.* Leipzig, 584 pages.

Shaw, E. W., and Trowbridge, A. C. (1916) *Description of the Galena and Elizabeth quadrangles, Iowa, Illinois,* U. S. Geol. Survey, Geol. Atlas, folio 200, 13 pages, maps.

Sutton, A. H. (1935) *Stratigraphy of the Silurian system of the upper Mississippi valley,* Kan. Geol. Soc., Guidebook 9th Ann. Field Conf. p. 268-280.

Thwaites, F. T. (1928) *Pre-Wisconsin terraces of the Driftless Area of Wisconsin,* Geol. Soc. Am., Bull., vol. 39, p. 621-642.

——— (1931) *Buried pre-Cambrian of Wisconsin,* Geol. Soc. Am., Bull., vol. 42, p. 719-750.

——— (1935) *Road log for fourth day,* Kan. Geol. Soc., Guidebook 9th Ann. Field Conf. p. 105-126.

Thwaites, F. T., and Twenhofel, W. H. (1920) *Windrow formation; an upland gravel formation of the Driftless and adjacent areas of the upper Mississippi valley,* Geol. Soc. Am., Bull., vol. 32, p. 293-314.

Trowbridge, A. C. (1921) *The erosional history of the Driftless area,* Univ. Iowa Studies, 1st ser., no. 40, Studies in Nat. Hist. vol. 9, no. 3, 127 p.

——— (1934) *Upper Mississippi Valley structure,* Geol. Soc. Am., Bull., vol. 45, p. 519-528.

——— (1935) *Kansas Geological Society Guidebook 9th Annual Field Conference, Upper Mississippi Valley.*

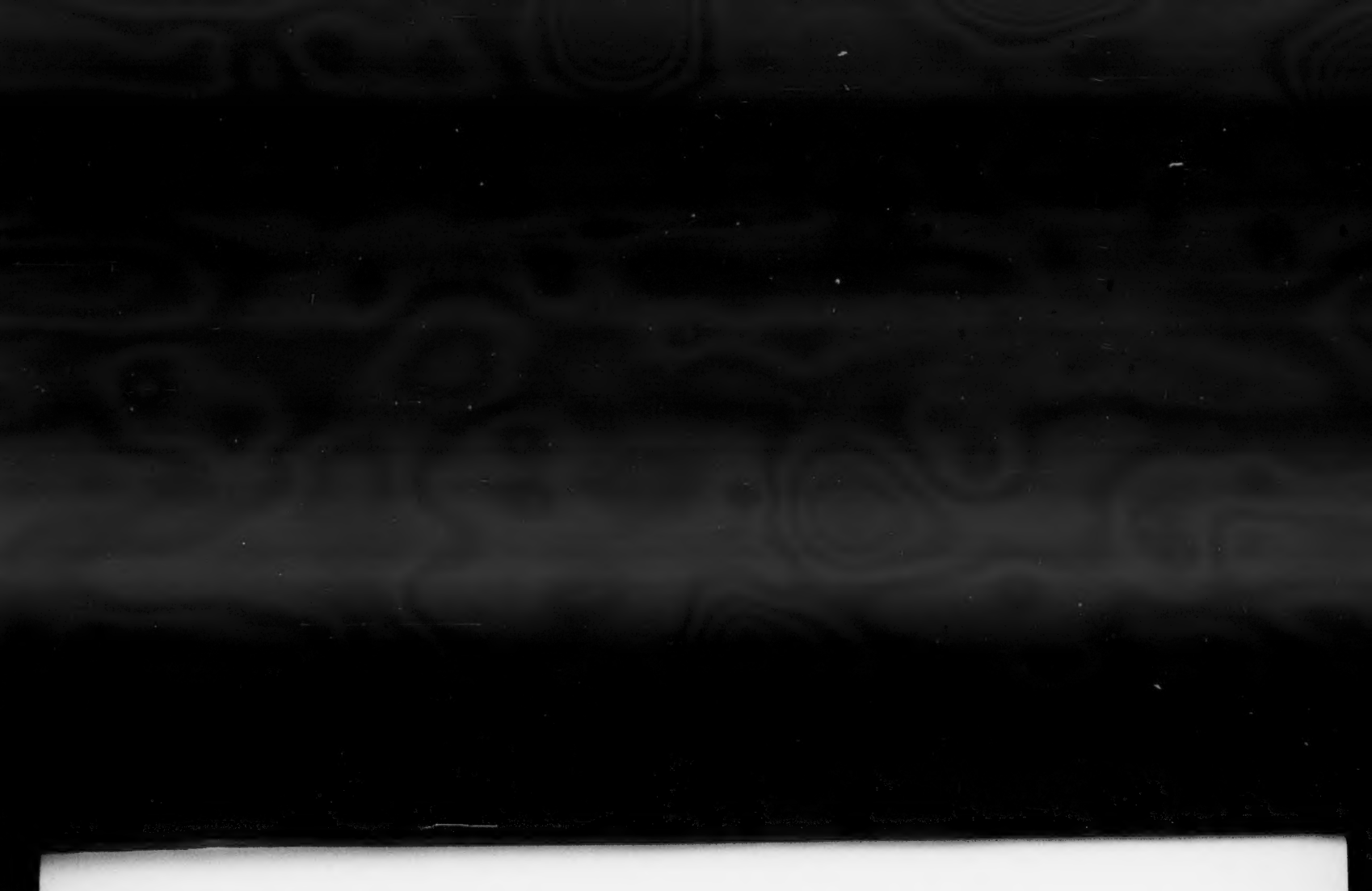
Van Hise, C. R. (1896) *A central Wisconsin base level,* Science, n.s., vol. 4, p. 57-59.

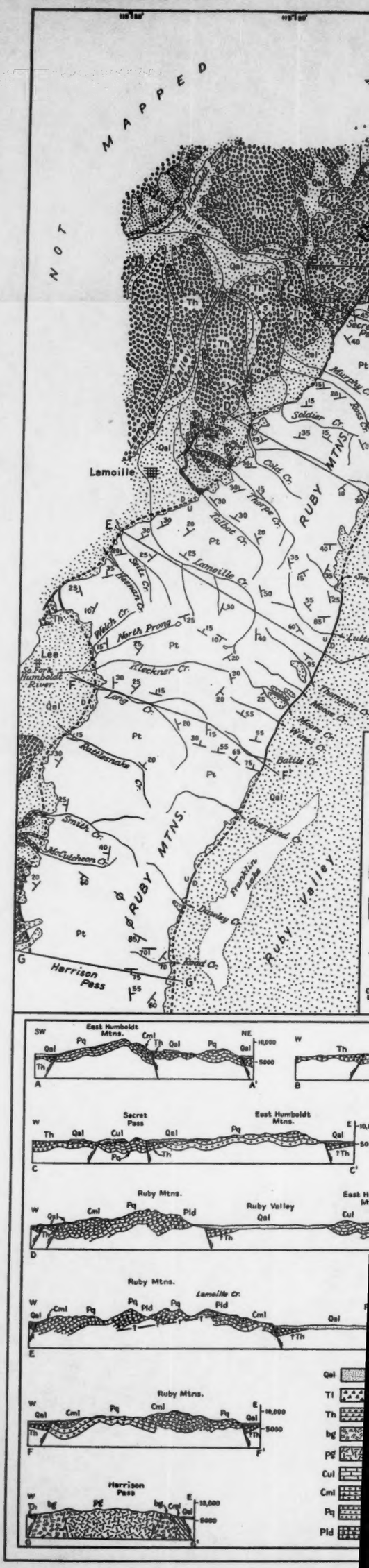
Weidman, Samuel (1903) *A pre-Potsdam peneplane of the pre-Cambrian of north-central Wisconsin,* Jour. Geol., vol. 11, p. 289-313.

Winslow, Arthur (1893) *The Osage river and its meanders,* Science, vol. 22, p. 31-32.

Western Reserve University, Cleveland, Ohio.

Manuscript received by the Secretary of the Society, May 16, 1938.





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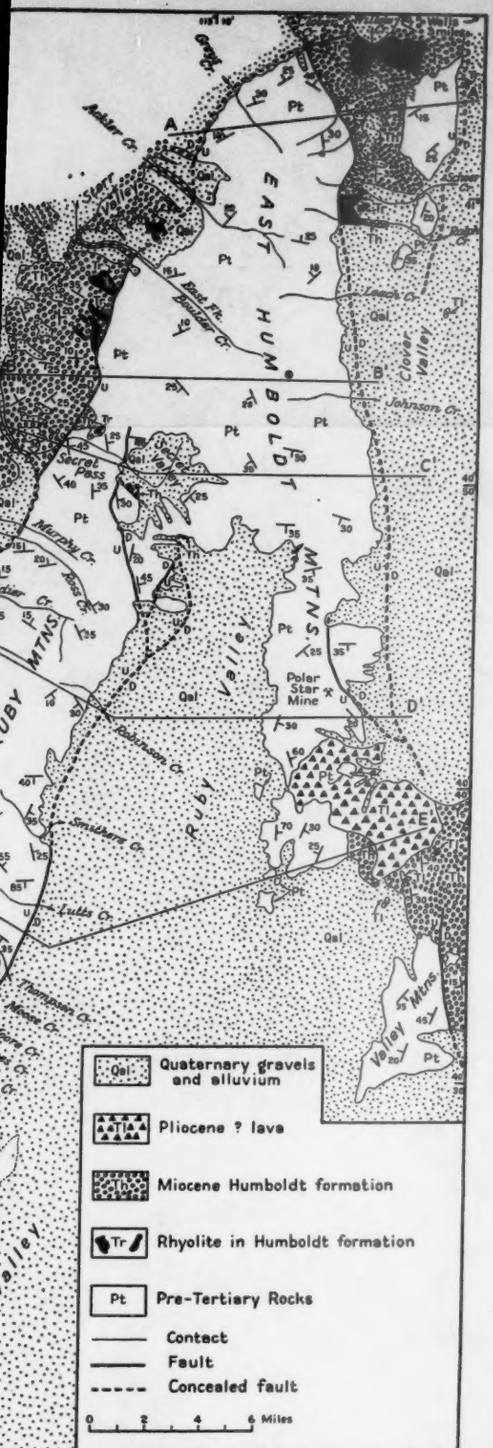

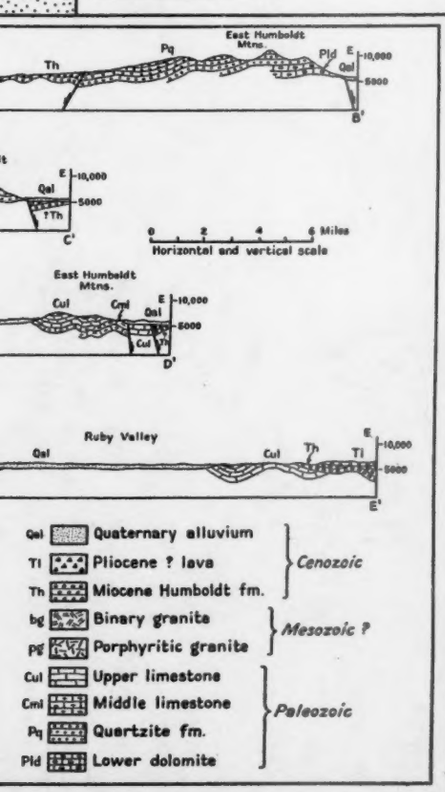

TIONS OF RUBY-EAST HUMBOLDT RANGE

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BASIN-RANGE STRUCTURE OF THE RUBY-EAST HUMBOLDT RANGE, NORTHEASTERN NEVADA

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ABSTRACT

The Ruby-East Humboldt Range consists of pre-Miocene igneous, metamorphic, and sedimentary rocks of complex structure. The adjoining basins contain deformed beds of the upper Miocene Humboldt formation. The boundary structure of the mountain block is well exposed because of dissection by the through-flowing Humboldt River.

This range is a westward-tilted horst bounded by normal faults which dip 60 to 70 degrees basinward. Displacements on the east boundary faults have been at least twice as great (5500 to 6000 feet) as on the west boundary faults (2000 feet). The northern termination of the mountains is due to intersection of the east and west boundary faults. The structure of the pre-Miocene rocks is discordant with the trend and shape of the range.

Five periods of Basin-Range faulting have been established: (1) Late middle or early upper Miocene, displacement small (open to question); (2) upper Miocene during deposition of the Humboldt formation, displacement larger; (3) later upper Miocene to Pliocene, younger than the Humboldt formation and older than the Pliocene (?) lava, amount of displacement unknown; (4) Pliocene to Pleistocene, later than the Pliocene (?) lava and extending to middle or late Pleistocene, the period of last major uplift of the range, displacement large; (5) late Pleistocene to Recent, later than the earliest Wisconsin, displacement small.

Lack of evidence of intense late Cenozoic compression indicates that the uplift is to be attributed to vertical forces operating while the crust was in a neutral or tensional state.

The scarps bounding the range are composite, in places rejuvenated by late Pleistocene or Recent fault movements.

INTRODUCTION

GENERAL STATEMENT

The Ruby-East Humboldt Range is an isolated Basin Range approximately midway between the eastern border (Wasatch Mountains) and the western border (Sierra Nevada) of the Great Basin of western United

States (Fig. 1). At the borders of the Basin-Range province the pure elements of Basin-Range structure are likely to be obscured and complicated by structures which overlap from neighboring provinces. It is by the study of ranges some distance within the province that we may even-

FIGURE 1.—*Index map*
Ruby-East Humboldt Range shown in solid black.

tually hope to obtain a true picture of the interrelated elements of Basin-Range structure. This paper presents structural, stratigraphic, and geomorphic data on the internal and boundary structure of a typical range located far within the boundaries of the Basin-Range province and having unusually favorable exposures of essential features.

The writer is greatly indebted to Professors Kirk Bryan and Marland P. Billings of Harvard University for invaluable suggestions and criticisms both in the field and in the office. Mr. Edward A. Schmitz has aided in the preparation of illustrations. Many residents of Nevada, too numerous to mention individually, have extended courtesies and kindness in the course of the field work.

LOCATION AND PHYSICAL FEATURES

The Ruby-East Humboldt Range [1] is located in Elko County, northeastern Nevada, 180 miles west of Salt Lake City, Utah. Latitude 40° 30′ north and longitude 115° 30′ west intersect near its center (Fig. 2).

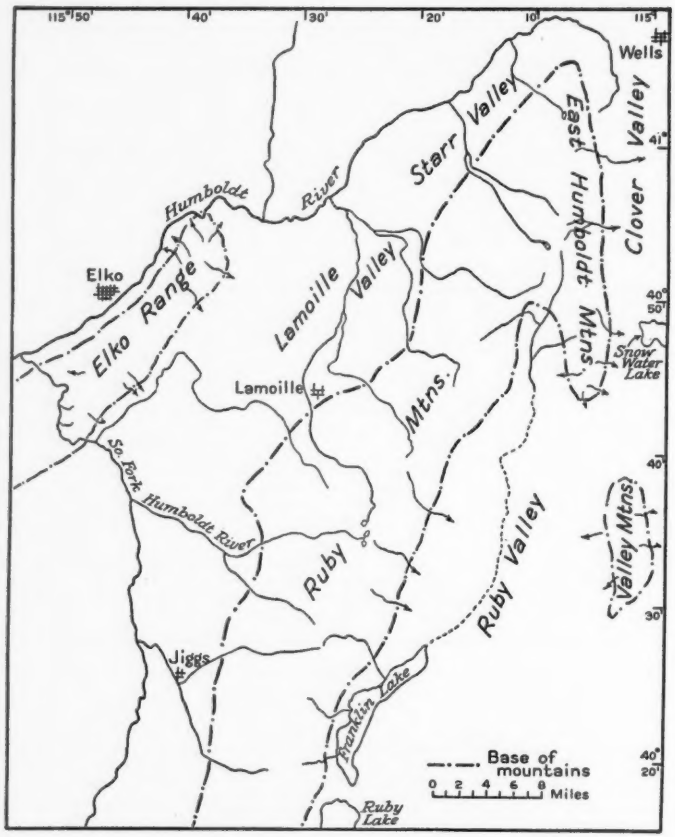

FIGURE 2.—*Geographic map*

Showing broad geographic features of the Ruby-East Humboldt region.

The range is divided by Secret Pass into a northern part—the East Humboldt Mountains—and a southern part—the Ruby Mountains. From an abrupt northern termination in a broad pass near Wells, it extends S. 20° W. for 100 miles. The mountain block has an average width of

[1] Not to be confused with the Humboldt Range of west-central Nevada.

8 to 9 miles and is bounded on the east and west by wide, relatively flat basins. The basins to the east are hydrographically closed and contain dry or intermittent lakes; those to the west are drained and dissected by the Humboldt River and its tributaries. Figure 2 shows these geographical relations and the location of the Elko Range to the west.

Throughout most of its length the Ruby-East Humboldt Range is asymmetric; the eastern face is steeper and more rugged than the western, and the crest lies east of the center of the range. The altitude of the crest averages about 10,000 feet, although in places it is higher, for just south of Lamoille Canyon seven peaks rise to altitudes between 11,300 and 11,400 feet (Pl. 2, fig. 1).

The larger part of the Ruby-East Humboldt Range is shown on the topographic sheets of the Halleck and Jiggs quadrangles, published in 1935 and 1937 respectively. These maps have been supplemented by special Forest Service maps kindly furnished by Mr. A. R. Torgerson of the Forest Service at Elko.

Five months were spent in the field during the summers of 1936 and 1937 mapping and studying the geology of the Ruby-East Humboldt Range and the adjoining basins. Only the northern half of the range, from Wells at the north end to Harrison Pass at the south, a total distance of 57 miles, has been studied.

PREVIOUS WORK

The first geological study of this region worthy of note was made by King and his associates (1877-1878). Later, the Wheeler Survey (1889) published geological observations on the region, and Spurr (1901, 1903) described briefly the southern part of the range. Hill (1916) published a rough reconnaissance map of part of the range showing the distribution of rock types, and Blackwelder (1931; 1934) described some of the glacial features.

A review of the history of work and thought and the whole mass of geological literature of the Great Basin and the Basin Ranges would be too lengthy for the present paper. The well-informed geologist is already familiar with the summaries of Gilbert (1928, p. 1-9) and Davis (1925, p. 387-388; 1930, p. 293-300). He will recognize that the landmarks in development of thought on these problems are the papers of Gilbert (1874; 1875; 1890; 1928), King (1878), Dutton (1880), I. C. Russell (1882; 1885; 1887), LeConte (1889), Spurr (1901), Davis (1903; 1905; 1913; 1925; 1930), Louderback (1904; 1907-1908; 1923; 1924), Lindgren (1911), R. J. Russell (1928), and Gilluly (1928; 1932).

It seems worth while to call attention to an observation concerning Basin-Range structure made in 1859 by Henry Englemann, a geologist

with the Simpson Expedition (Simpson, 1876) across the Great Basin. Englemann's statement, which follows, has passed unnoticed, even by Gilbert, and his contribution has been unrecognized.

"The whole [the Great Basin] must be regarded not as composed of separate mountain chains, but as one system, one great continental swell, the relief of which has been shaped by numerous parallel fissures and corresponding mountainous upheavals running nearly north-south."

He writes further of sporadic upheavals separated by intervals of comparative rest and states that the mountains were not all formed by one violent effort, but that considerable time was involved in their formation. This description is strikingly similar to present-day conceptions of Basin-Range structure.

LITHOLOGY AND STRATIGRAPHY

GENERAL STATEMENT

The rocks of the Ruby-East Humboldt Range and the adjoining basins can be divided into two groups: (1) Pre-Miocene group, composing the mountain block, (2) late Cenozoic group, found chiefly in the adjoining basins. The columnar section (Fig. 3) shows the relations, nature, age, and thicknesses of the rock units. The pre-Miocene group has been studied in reconnaissance fashion, and the division of these rocks into formations is tentative. Dr. George H. Anderson of the California Institute of Technology is to make a detailed study of these rocks, so only a bare outline of their chief characteristics is given.

PRE-MIOCENE ROCKS OF THE MOUNTAIN BLOCK

General description.—The Ruby-East Humboldt Range is made up of a complex of sedimentary and metamorphic rocks intimately injected by dikes, sills, and irregular igneous bodies (Pl. 2, fig. 2). In many places 30 to 40 per cent, and, in a few places, all of the rocks exposed are igneous. In general the degree of metamorphism is greater in the older rocks, though there are many exceptions. Geographical names have not been used for the formations distinguished within the pre-Miocene rocks in order to avoid setting a precedent for the studies of Dr. Anderson. The sedimentary and metamorphic rocks are all thought to be Paleozoic, not in part pre-Cambrian as suggested by King (1878, p. 12, 62; 1877, p. 528, 537) and more recently by Blackwelder (1935, p. 157).

Lower dolomite.—The Lower dolomite is the oldest formation and consists of sillimanite-garnet schist, quartz-biotite schist, quartzite, marble, and diopside granulite.[2] It has been intimately intruded by dikes, sills, and irregular bodies of pegmatite and granitic rocks, so that in most places

[2] Granulite is used in the sense defined by Holmes (1928, p. 111): "A granulose metamorphic rock, composed of even-grained interlocking granular minerals."

FIGURE 1. RUGGED TOPOGRAPHY AND ROCK EXPOSURES
Looking north near head of Kleckner Creek.

FIGURE 2. CREST OF RANGE SOUTH FROM LUTTS CREEK
Rocks are complexly folded marble, schist, and lime-silicate rocks intruded by pegmatite and binary granite.

FIGURE 1. EAST FACE WEST OF CLOVER VALLEY
Note gently dipping beds appearing as horizontal bands near crest of range.

FIGURE 2. GENTLY DIPPING BEDS OF EAST FACE
Near head of Leach Creek.

EAST SIDE OF EAST HUMBOLDT MOUNTAINS

30 to 60 per cent of the rocks exposed are igneous. Including the intruded igneous rocks, the thickness of this formation is conservatively estimated at 5000 feet. It is believed to be Paleozoic, as it is part of a conformable series in which overlying beds are Carboniferous (King, 1878, p. 193, 205).

AGE	FORMATION	SYMBOL	COLUMNAR SECTION	LITHOLOGY	THICKNESS in feet
QUATERNARY	DRIFT, GRAVELS, ALLUVIUM	Qal		Pediment and terrace gravels, glacial drift, recent alluvium.	5-150
PLIOCENE ?	PLIOCENE ? LAVA	Tl		Pyroxene andesite and olivine-pyroxene basalt flows, breccia, and tuff.	200-400
	UNCONFORMITY				
MIOCENE	HUMBOLDT FORMATION	Th		Breccia, fanglomerate, conglomerate, sandstone, mudstone, shale, oil shale, limestone, ash, tuff, and interbedded rhyolite flows (Tr)	5800
	RHYOLITE	Tr			
	UNCONFORMITY				
EARLY TERTIARY OR LATE MESOZOIC	BINARY GRANITE PORPHYRITIC GRANITE	bg pg		Binary granite, porphyritic granite, pegmatite.	
CARBONIFEROUS	UPPER LIMESTONE	Cul		Gray, black, buff, thin-bedded limestone, fossiliferous	5000 + ?
	MIDDLE LIMESTONE	Cml		Massive gray, white, and buff limestone, fossils scarce.	4000 ± 1000
CARBONIFEROUS OR OLDER PALEOZOIC	QUARTZITE FORMATION	Pq		Quartzite, quartz-mica schist, marble, biotite gneiss, locally conglomerate and breccia	4000 ±
	LOWER DOLOMITE	Pd		Diopside granulite, quartz-mica schist, sillimanite-garnet schist, quartzite, and marble.	5000 + ?

FIGURE 3.—*Columnar section*

Quartzite formation.—The Quartzite formation conformably overlies the Lower dolomite. It consists largely of brown quartzite and smaller amounts of quartz-biotite schist, marble, granulite, and biotite gneiss. This formation is also intimately intruded by igneous rocks but not to so great an extent as the Lower dolomite. Calculations in areas of uniform structure give a thickness, including igneous intrusions, ranging from 2000 to 4000 feet. It is conformable with overlying beds which contain Carboniferous fossils and is, therefore, thought to be Paleozoic.

Middle limestone.—The Middle limestone conformably overlies the Quartzite formation and is composed of gray, brown, buff, and white limestone which contains minor amounts of arenaceous and argillaceous material. In places it has been metamorphosed to marble, quartzite, quartz-mica schist, and diopside granulite. Numerous pods, dikes, and

sills of pegmatite and granitic rocks intrude the formation. The Middle limestone, including igneous rocks, is about 4000 feet thick. It is from this formation that King (1878, p. 205) collected Carboniferous fossils.

Upper limestone.—The Upper limestone appears to overlie the Middle limestone with a slight and perhaps local disconformity. The Upper limestone is composed of moderately thin beds of buff, brown, gray, and black limestone, some arenaceous and argillaceous limestone, and a few chert-pebble conglomerate beds. The Upper limestone is cut by dikes, sills, and pods of pegmatite and granitic rocks, but to a less extent than are the underlying formations. A conservative estimate of the thickness, including intruded igneous rocks, is 5000 feet. Hill (1916, p. 56) collected some fossils, identified as probably Permian, from beds in this formation.

Igneous rocks.—Dikes, sills, and irregular masses of granite and pegmatite have been intruded into the deformed pre-Miocene rocks. The amount of igneous rock increases more or less progressively from the north end of the range southward to Harrison Pass. South of Harrison Pass the amount of igneous material decreases sharply, and the mountains are composed chiefly of sedimentary rocks. In the central part of the range the igneous intrusions make up a network of crisscrossing dikes and sills of medium-grained binary granite and coarse-grained pegmatite. At Harrison Pass is a stock-like body, 7 miles across, of porphyritic granite. The intrusions may be Late Jurassic or early Tertiary.

LATE CENOZOIC ROCKS

Miocene Humboldt formation.—Continental deposits belonging to the Humboldt formation lie in basins adjoining the Ruby-East Humboldt Range and have been described in another paper (Sharp, 1939). Only a brief résumé of their features will be given here. The Humboldt formation is composed of fluviatile and lacustrine deposits, mostly conglomerate, sandstone, siltstone, mudstone, and shale, with minor amounts of freshwater limestone, limy shale, oil shale, diatomite, ash, tuff, fanglomerate, and breccia. The formation contains a few thin interbedded rhyolite flows.

The older rocks of the mountains have supplied detritus to these basin deposits. Breccias and fanglomerates derived from the actively growing fault scarp of the East Humboldt Mountains are interbedded in the Humboldt.

Beds of the Humboldt formation are either unconformable on or in fault contact with the older rocks. The Humboldt beds have a simple structure and gentle dips in the center of the basins and steeper dips and more complex structure near the edges of the mountain block. The forma-

tion has a measured thickness of 5800 feet and may be somewhat thicker. Paleontological evidence shows that it is upper Miocene (Sharp, 1939).

Pliocene (?) lava.—Flows, breccias, and tuff, 200 to 400 feet thick, composed of pyroxene andesite and pyroxene-olivine basalt rest unconformably on the Humboldt and pre-Miocene formations. These volcanic rocks are shown on the map (Pl. 1) at the south end of the East Humboldt Mountains. Another smaller area of the same lavas has been studied in the Elko Range near Elko (Fig. 2). The Pliocene (?) lavas rest upon an erosion surface of low relief which truncates deformed upper Miocene and pre-Miocene rocks. A period of deformation and erosion ensued between deposition of the upper Miocene Humboldt formation and extrusion of the lavas. For that reason, these lavas are thought to be Pliocene (?). They appear too old to be Pleistocene. Similar relations between andesitic and basaltic lavas and deformed Miocene deposits have been reported from other parts of the Great Basin by Louderback (1904; 1907-1908; 1924), Buwalda (1914), and others.

Quaternary deposits.—Quaternary gravel covers terraces and pediments on the flanks of the Ruby-East Humboldt Range. It is mostly stream gravel and slope-wash detritus, composed of subangular to rounded pebbles, cobbles, and boulders of quartzite, gneiss, limestone, and granitic rocks, imbedded in a sandy to gravelly matrix. In so far as observed, these deposits never exceed 25 feet in thickness, and they overlie unconformably the deformed and eroded Miocene basin deposits and the older rocks of the mountain block.

Drift of two substages of the Wisconsin glaciation has been deposited in the mountains and in a few places on the piedmont slopes. In Figure 3 of Plate 4, the long smooth bare ridges (m) descending to the base of the range are lateral moraines of the earliest Wisconsin glaciation. Inside of these moraines may be seen the smaller lateral and terminal moraines of the later Wisconsin glaciation. These deposits and the related features of glaciation are described in another paper (Sharp, 1938).

INTERNAL STRUCTURE OF THE RANGE

GENERAL STATEMENT

The internal structure of the range has been studied in reconnaissance, but the entire areal distribution of the various pre-Miocene rock units has not been mapped. However, reasonably detailed, east-west sections showing the relations of these units have been made across the range at intervals from north to south. These sections and the map (Pl. 1) illustrate something of the internal structures and show that they are discordant with the trend and shape of the range. Minor igneous intrusions

have been omitted in the sections, although, from Secret Pass southward, igneous rocks make up 30 to 60 per cent of the rocks exposed. The original structure has been restored by data gathered from patches, remnants, or inclusions of the older rocks in the igneous intrusions.

FOLDS

Only a few major folds are described here as examples. From Lamoille Creek to Soldier Creek the west face of the Ruby Mountains is made up of beds on the northwest limb of an anticline which plunges 15 to 30 degrees northeastward. Near Soldier Creek this fold swings sharply to the northwest and is truncated by the west face of the mountains. Beds in Secret Pass, farther north, dip eastward and are on the east limb of this anticline. The west slope of the East Humboldt Mountains north of Secret Pass is made up of beds on the west limb of an anticline which lies northeast of the one just described. The axis of this second anticline trends nearly due north, and beds on its east limb are exposed on the east side of the range at the north end. North of Johnson Creek this fold plunges 10 to 25 degrees northward; south of Johnson Creek it plunges 25 to 35 degrees southward. Gently dipping beds near the crest of this fold are exposed in the east face of the central East Humboldt Mountains (Pl. 3, fig. 2) showing that the eastern limb of the fold is cut off by the face of the range.

South of Lamoille Canyon the internal structure is more complex. Parts of two anticlines trending roughly west-northwest and strongly asymmetric to the southwest have been mapped along Battle Creek and Road Canyon on the east side of the range. These folds are truncated at a large angle by the east face of the range. The complexity and diverse trends of the structures in the southern part of the range may be related to the igneous intrusions so abundant there.

The folding occurred after deposition of the Permian (?) Upper limestone and before or contemporaneously with intrusion of the Late Jurassic or early Tertiary igneous rocks.

FAULTS

Some normal faults have been mapped or observed within the range, particularly on the east side of the Ruby Mountains near the north end of Ruby Valley and on the east side of the East Humboldt Mountains near the Polar Star mine. These faults involve only pre-Miocene rocks and trend obliquely to the front of the range. They are later than the igneous intrusions, hence at least post-Jurassic, but, as they have no direct topographic expression, they are thought to be older than the faults which bound the range.

In Secret Pass, beds of the Upper limestone have been thrust over

igneous intrusives and quartzite, gneiss, and schist of the Quartzite formation. The slickensided surface of the fault plane was observed at one locality. Here the plane is essentially horizontal, striations trend N. 70° E., and the minor features of the slickensided surface suggest thrusting from the west. The thrust fault cuts igneous intrusive bodies and for that reason must be post-Jurassic. It is older than the Miocene basin deposits, and by analogy with other parts of Nevada and western Utah (Eardley, 1934; Ferguson, 1929; Hewett, 1926; 1928a; 1931; Longwell, 1921; 1928; Nolan, 1935; Westgate, 1932) the thrusting may be tentatively dated as early Tertiary. This thrust was mistaken by King (1878, p. 65) for an unconformity.

In summary, the trend and form of the Ruby-East Humboldt Range shows little relation to the internal structure. This discordance is best shown on the east side where the folds of the mountain block are sharply cut off by the steep eastern face of the range. In places the western side of the range is almost a dip slope, and a closer relation exists between the internal structure and the trend of the front of the range. However, in a generalized way, the internal structure is truncated by the west front also.

BOUNDARY STRUCTURE

GENERAL STATEMENT

Many ranges of the Great Basin are bounded on both sides by closed, detritus-filled basins. The detrital filling of the basins has, in most places, buried the flanks of the mountain block making direct study of the boundary structure difficult, if not impossible. The Ruby-East Humboldt Range is one of the rare examples of a Basin Range, bounded on one side—the west—by a dissected basin with through-flowing drainage in which the boundary structure is exposed, and on the other side—the east—by a closed basin filled with undissected detritus. This relation offers the opportunity of applying geologic and geomorphic criteria of faulting, developed in a dissected basin, where the structural evidence of faulting can be directly observed, to a closed and filled basin where the boundary structure itself cannot be directly observed.

Geologic criteria of faulting which proved of use in this study were: actual exposures of the fault plane or zone, drag, brecciation, minor faults, juxtaposition of different formations where the relations cannot be explained by folding or deposition, truncation of beds or structures, coarse detrital deposits such as breccias and fanglomerates derived from a growing fault scarp, and hot and cold springs. Geomorphic criteria of faulting which proved of use were: triangular facets, piedmont scarps, dissected pediments on the mountain block where flanked by a closed basin, and a linear range front without projecting spurs.

DESCRIPTION OF BOUNDARY FAULTS

Faults on east side of East Humboldt Mountains.—The faults at the east base of the range are covered by Quaternary basin-fill except at the northern end, so little direct geologic evidence of faulting is available. As shown in Plate 3, the east front of the East Humboldt Mountains is a remarkably abrupt, steep scarp rising 4000 to 5000 feet above the floor of Clover Valley. This scarp is one of the finest in the Great Basin and rivals the east face of the Sierra Nevada in ruggedness and beauty, though it is not so high.

At the north end of the East Humboldt Mountains a subsidiary westward-tilted fault block, bounded by a fault on the east side, lies east of the base of the range. Between the base of the range and this forelying block, Miocene Humboldt beds have been squeezed into an open syncline by uplift and tilting of the blocks. These Miocene beds are separated from the pre-Miocene rocks of the East Humboldt Mountains by the east boundary fault which has been observed on the northern bank of Willow Creek, NW¼ sec. 2, T. 36 N., R. 61 E. At this place the fault dips 73 degrees east, or basinward, and it is easily traced 4 miles north-northwestward from Willow Creek to the end of the range. The Miocene beds exposed on the east side of the fault throughout this distance dip as steeply as 50 degrees eastward owing to drag along the fault. They are also cut by a number of minor basinward-dipping normal faults which roughly parallel the principal boundary fault. Coarse breccias and fanglomerates interbedded in the Humboldt formation indicate an earlier (late Miocene) movement on this boundary fault. Since the boundary fault dips eastward, the direction of drag and the position of the younger beds on the block to the east show that it is a normal fault.

At the mouths of Willow and Clover creeks, moraines of the earliest and later Wisconsin glaciations lie across the fault and are unbroken. Displacements on the boundary faults will be discussed at greater length in a succeeding section.

The evidence that the subsidiary block, already described, is faulted on the east side is largely geomorphic, although a small patch of rhyolite, of the type interbedded in the Miocene deposits, lies immediately east of the block and suggests faulting. This subsidiary fault, 4 miles east of the principal boundary fault, dies out southward toward Clover Valley within a distance of 9 to 10 miles from its northern end.

The fault at the foot of the principal scarp of the East Humboldt Mountains continues for 25 miles nearly due south from Clover Creek along the base of the range. South of Ralph Creek it is buried by Quaternary fan deposits, and the only geologic evidence of faulting is the relation of the internal structure and the range front. The internal

FIGURE 1. EAST FACE OF NORTHERN RUBY MOUNTAINS
This precipitous scarp is 3000 feet high.

FIGURE 2. PIEDMONT SCARP NEAR LUTTS CREEK
This scarp, composed of pre-Miocene igneous intrusives and Paleozoic limestone, is 200 feet high.

FIGURE 3. WEST FACE OF RUBY MOUNTAINS NEAR LAMOILLE
Smooth ridges (m) descending to the foot of the range are earliest Wisconsin glacial moraines.

FIGURE 1. PIEDMONT SCARP AT GREYS CREEK
The scarp in the mid-foreground is 75 feet high and cuts fan gravels.

FIGURE 2. GRABEN-LIKE DEPRESSION
Looking northward along the base of the piedmont scarp shown in Figure 1.

FIGURE 3. PIEDMONT SCARP AT SEITZ CREEK
Earliest Wisconsin glacial moraines (m) cut by the low piedmont scarp which extends along the base of the range.

PIEDMONT SCARPS AT WEST BASE OF RANGE

structure of the East Humboldt Mountains is a broad anticline, trending nearly north-south; the nose can be seen at the north end of the range. A large part of the east limb of this fold is missing south of Angel Creek. The beds exposed in the east face of the East Humboldt Mountains, west of Clover Valley, are horizontal or gently dipping, as shown in Plate 3. The fold is bisected by the range front, and the eastern limb has been dropped from view. To the north and south of Clover Valley, the fold axis and the range front diverge so that a larger part of the eastern limb of the fold is preserved. The linear trend of the Clover Valley scarp, its abruptness, and the lack of projecting spurs all strongly suggest origin by faulting. The vertical component of the dip-slip displacement on the boundary fault in Clover Valley is 4500 to 5800 feet at minimum, judging solely from the topographic relief. Southward the height of the East Humboldt Mountains decreases, and the boundary fault appears to be dying out. At the Polar Star mine the displacement on the fault appears to be 2000 to 2500 feet. Just east of the Polar Star mine is a subsidiary fault which has no topographic expression and is older than the boundary fault.

Eleven miles farther south a fault bounds the Valley Mountains on the east side; along this fault westward-dipping fresh-water limestone and ash beds of the Humboldt formation have been brought into contact with the Carboniferous limestone of the Valley Mountains. This fault may be a southward continuation of the fault at the east base of the East Humboldt Mountains, but no evidence for connecting the two was found in the field.

Faults on east side of Ruby Mountains.—The south end of the East Humboldt Mountains extends as a southerly projecting ridge a few miles east of the north part of the Ruby Mountains. The north end of Ruby Valley forms a re-entrant between the Ruby and East Humboldt mountains, so that the northern 7 or 8 miles of the east front of the Ruby Mountains faces the back slope of the southern part of the East Humboldt block. Farther south, the eastern face of the Ruby Mountains is more clearly exposed and is as magnificent as the East Humboldt scarp near Clover Valley (Pl. 4, fig. 1).

Geologic and geomorphic features show that the Ruby Mountains are bordered by a fault along their east base. From Battle Creek southward the axes of the folds of the mountain block are oblique to the east face, in places at large angles. North of Battle Creek the fold axes are more nearly parallel to the trend of the range, but the limbs of these folds are cut off cleanly by the east face (section DD′, Pl. 1). Two small outcrops of fine-grained Tertiary beds at the north end of Ruby Valley

suggest that at least part of Ruby Valley may be underlain by Tertiary beds which have been downfaulted into contact with the older rocks of the mountain block.

A fault scarp 50 to 200 feet high and 10½ miles long breaks the piedmont slope between Smithers Creek and Battle Creek (Pl. 4, fig. 2).

FIGURE 4.—*Lutts scarp, east side of Ruby Mountains*
The scarp is separated from the base of the mountain slope by a flat pediment nearly a mile wide. (Sketched from a photograph)

Between this scarp and the base of the range a quarter of a mile to 1¼ miles to the west, a broad dissected pediment is cut on the pre-Miocene rocks of the mountain block (Fig. 4). Just north of Dawley Creek is a lower but similar scarp. These scarps indicate the position of the boundary fault and will be discussed at greater length in a later section of this paper.

Remnants of pediments on the mountain block have been noted near Overland Creek and other creeks to the south. The fact that these pediments are dissected by streams draining into a closed basin suggests that they have been differentially uplifted by faulting.

The features just described strengthen the argument that the linear trend and abruptness of the east face of the Ruby Mountains are the result of faulting. The east boundary fault of the Ruby Mountains extends from the northern end of Ruby Valley S. 25° E. for 35 miles to Harrison Pass. The fault has not been studied south of Harrison Pass, but reconnaissance shows that it extends southward for another 20 miles.

A minimum value for the vertical component of the greatest dip-slip displacement on the east boundary fault of the Ruby Mountains is between 4500 and 5000 feet. The fault appears to maintain about this displacement for 35 miles southward from a point 6 or 7 miles south of its northern end.

The fault at the east base of the Ruby Mountains may do one of three things at its northern end: (1) It may die out because of decreasing

displacement; (2) it may pass into the southern part of the East Humboldt block; (3) it may be terminated by a cross fault. The first of these explanations appears to be the correct one. Displacement on the fault decreases progressively northward from a point 6 miles south

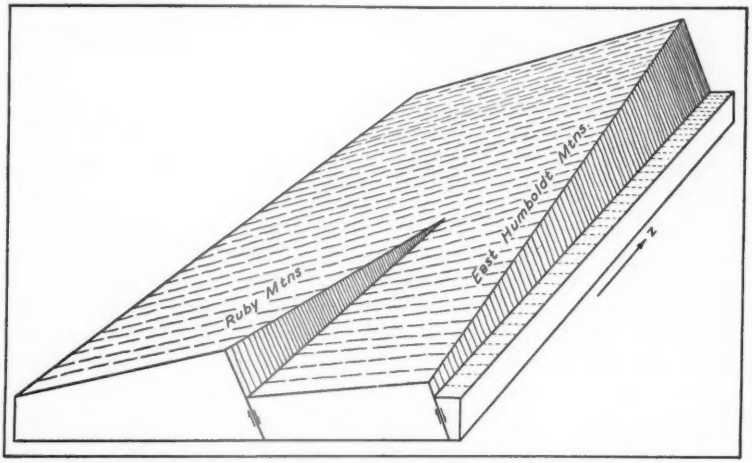

FIGURE 5.—*Block diagram of Ruby and East Humboldt blocks*
Illustrating en echelon arrangement and complementary displacement on faults. The block is about 15 miles wide and the vertical exaggeration is about twofold.

of Secret Pass. The termination of the fault cannot be seen, but detailed field studies by pace and compass mapping failed to show that this fault passes into the main mass of the East Humboldt Mountains, or that it is terminated by a cross fault. It appears to die out at its northern end with an average decrease in throw of 130 feet per 1000 feet along the strike. This ratio is only slightly greater than that reported by Lasky (1936, p. 53) from the Bayard District, New Mexico.

The fault at the east base of the East Humboldt Mountains, although parallel and only 8 miles to the east, increases in displacement northward. The displacements on these two faults are thus complementary. This complementary displacement plus westward tilting of both blocks appears to explain the en echelon relations of the southern East Humboldt Mountains and the northern Ruby Mountains. A rough block diagram, Figure 5, represents the interpretation offered. In the diagram, the pre-faulting surface is assumed to be essentially flat, and erosion during or after uplift is not considered. Both assumptions are incorrect, but their incorrectness does not invalidate the geometric relations of the

two mountain blocks. Faulting on the west side of the Ruby Mountains is not shown, as it is only a minor feature of the Ruby block and not here pertinent.

Secret Valley, a small intermontane basin just northwest of the north end of Ruby Valley, is a minor feature typical of Basin-Range structure. Secret Valley is underlain by a relatively downdropped block of Miocene beds, faulted on two sides and resting with depositional contact on the older rocks in other places.

Faults on west side of range.—The west face of the mountains is essentially linear despite local projections and re-entrants (Pl. 1). Geologic and geomorphic evidence for faulting at the west base of the range is good.

Just south of the north end of the range, remnants of a piedmont scarp [3] at the base of the mountains can be traced 3 miles south to Greys Creek (Pl. 5, fig. 1). Farther south, between Ackler Creek and Secret Creek, good exposures of Miocene beds give ample geologic proof of faulting. In a number of places fine-grained Miocene beds dip directly toward the steep scarp of the mountain block. That these fine-grained beds could be deposited so close to a steep scarp does not seem possible, and the attitude of the beds shows that the relations are due to faulting rather than folding. (*See* sections BB′ and CC′ of Plate 1.)

In Secret Pass a wide brecciated zone separates the pre-Miocene rocks of the mountain block from the basin deposits. A minor subsidiary fault is exposed in the basin deposits along the south bank of Secret Creek half a mile west of the boundary fault. These features further substantiate the presence of a fault at the west base of the mountains.

South of Secret Pass, remnants of a piedmont scarp extend along the base of the mountains for 2½ miles to the mouth of Ross Creek. This scarp appears to be composed entirely of fan deposits. Remnants of another piedmont scarp extend almost continuously from Talbot Creek south for 9 miles along the base of the range.

Isolated outcrops of Miocene beds along the mountain front south of Secret Pass and some well-preserved triangular facets at the mouths of Ross and Murphy creeks (Fig. 6) are further evidence of a fault at the west base of the Ruby Mountains in this section.

Evidence for faulting along the west front is poorer in the region of the South Fork of the Humboldt River and southward. Nevertheless, isolated outcrops of fine-grained basin deposits and the truncation of the internal structure of the range indicate a fault between the Miocene beds and the older rocks of the mountains. The displacement on the west

[3] Piedmont scarp is used in the sense defined by Gilbert (1928, p. 34) as a scarp on the piedmont slope caused by dislocation.

boundary fault, however, appears to be decreasing southward, and south of Harrison Pass the range may be simply a tilted block, faulted only on the east side.

The west boundary fault bifurcates at two places—at the mouth of Cold Creek and between Heenan and Welch creeks (Pl. 1). At these places

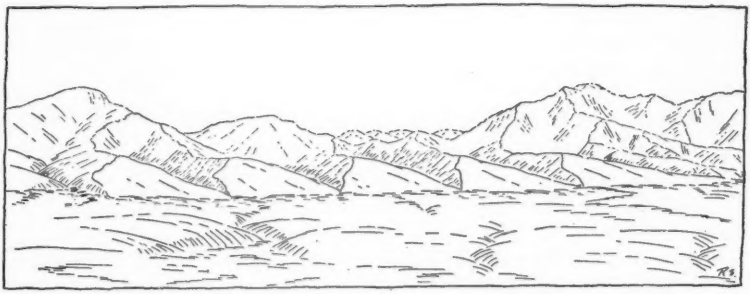

FIGURE 6.—*Triangular facets, west side of Ruby Mountains*
These facets slope about 25 degrees, which is much less than the dip of the fault plane. (Sketched from a photograph)

the principal fault swings sharply toward the mountain block for a short distance before resuming its general south-southwest trend. A minor fault splits off from the principal fault where it first swings mountainward and extends for a short distance along a line which is a continuation of the trend of the principal fault before bifurcation.

The evidence that the fault actually bifurcates is strongest between Heenan and Welch creeks where the trace of the subsidiary fault is marked by a piedmont scarp. The position of the main boundary fault is shown by a contact between the Upper limestone and the Quartzite formation, the intervening Middle limestone having been cut out by the fault. The evidence for bifurcation at the mouth of Cold Creek is not so strong, but even here the position of the subsidiary fault west of the principal fault is shown by a small upfaulted block of pre-Tertiary limestone on Thorpe Creek (Pl. 1). Each of these bifurcations is accompanied by a notable offset in the west base of the range, and it is possible that irregularities along the bases of other Basin Ranges may be related to a similar fault pattern.

The pre-faulting structure in the older rocks may have exerted an influence on the trend of the west boundary fault. Hulin (1930) has recently emphasized the influence of pre-faulting structures on the trend of faults bounding Basin Ranges, and similar relations have been noted by Eardley (1934, p. 389), Ferguson (1924, p. 377), Gilbert (1928, p. 21), and others.

NATURE OF BOUNDARY FAULTS

The boundary faults of the Ruby-East Humboldt Range are well enough exposed for measurements of the dip of the fault plane in only two localities:

(1) The plane of the boundary fault dips 73 degrees east or basinward on the east side of the East Humboldt Mountains near the north end on Willow Creek, NW¼ sec. 2, T. 36 N., R. 61 E.

(2) A fault between the Miocene beds and the pre-Miocene rocks of the mountains dips 60 degrees northeast or basinward in Secret Valley, NE¼ sec. 17, T. 34 N., R. 60 E.

At the west entrance to Secret Pass along the south bank of Secret Creek, eastern edge of sec. 2, T. 34 N., R. 59 E., a subsidiary fault, half a mile west of the principal boundary fault, is well exposed. The fault plane strikes N. 35° W. and dips 50 degrees basinward. This is a normal fault as the displacement has brought Quaternary alluvium on the west side into contact with Miocene basin deposits (Fig. 7), and it probably reflects the nature of the boundary fault.

Where exposed the boundary faults are normal faults. Minor normal faults have been observed within the basin deposits at several localities within and outside of the area shown on the map (Pl. 1). No evidence of thrust faults involving the Miocene basin deposits has been observed. The Miocene beds are deformed largely by simple tilting and warping, not by compression sufficiently intensive to imply compressional forces in the crust.

NATURE OF DISPLACEMENTS ON BOUNDARY FAULTS

The general assumption has been that displacement on the boundary faults of Basin Ranges has been largely, if not entirely, dip slip. Such seems to have been the case in the Ruby-East Humboldt Range, although some of the latest displacements on the faults may have had a strike-slip component.

Oblique striations have been observed on the plane of a fault at the mouth of Secret Pass (Fig. 7). This fault strikes N. 35° W. and dips 50° SW, striations on the fault plane pitch northwestward. The inclination of the striations indicates that the last movement of the mountain block has been relatively upward and southward. Striated fault surfaces in Secret Valley and at the north end of the range along Willow and Clover creeks show that at least some of the movement in these places was essentially a dip-slip displacement.

Evidence of at least some recent horizontal displacement on Basin-Range faults has been reported by R. J. Russell (1928, p. 454), Gianella and Callaghan (1934, p. 2, 20), Hobbs (1910, p. 379), Callaghan and Gianella (1935, p. 167), Lawson (1912, p. 197), and Pack (1926, p. 405).

Some of the latest fault movements in the Great Basin have been oblique and not simply dip slip. The total amount of strike-slip displacement on Basin-Range faults is not known. King (1878, p. 742-743) suggested horizontal displacements of as much as 5 miles on Basin-Range faults,

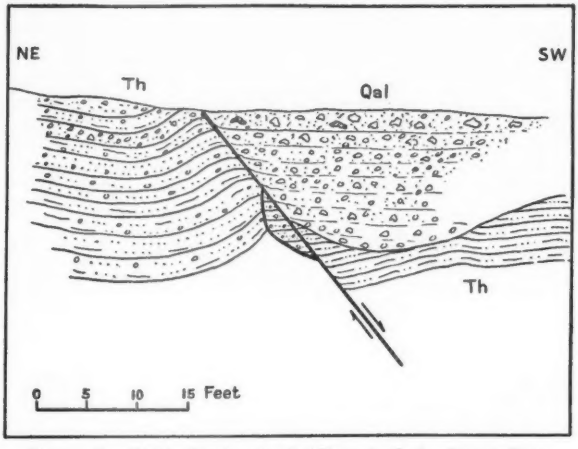

FIGURE 7.—*Field sketch of subsidiary fault in Secret Pass*
Qal—Quaternary alluvium; *Th*—Miocene Humboldt formation.

but his evidence is not convincing. No good evidence of extensive horizontal displacements on the boundary faults of the Ruby-East Humboldt Range was found, and it is difficult to imagine much horizontal displacement along irregular range bases such as the west base of the Ruby Mountains or the east base of the Sierra Nevada as noted by Davis (1927, p. 58).

There is some evidence of a slight absolute upward movement of the Ruby Mountain block, but the argument is too lengthy and indecisive to present here. The lack of glaciation earlier than the Wisconsin in this range may indicate a considerable absolute uplift of the mountain block in the late Pleistocene, but further evidence is needed on this point.

AMOUNT OF DISPLACEMENT ON BOUNDARY FAULTS

As shown in Figure 8 the difference in altitude between the crest of the range and the base of the Miocene deposits on the east side of the range gives directly a minimum value for the vertical component of the dip-slip displacement. The true displacement is greater by the thickness of rock which has been removed by erosion from the crest of the range since faulting began. On the west side of the range, the vertical component of the dip slip is represented by the thickness of the Miocene deposits imme-

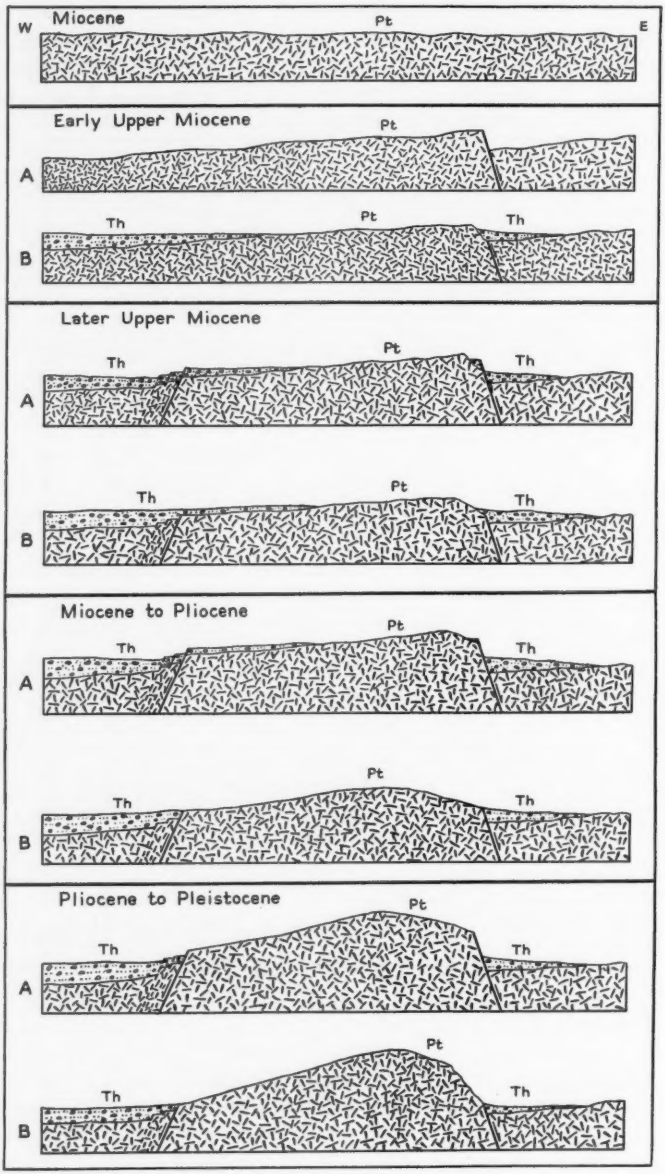

FIGURE 8.—*Sections showing fault-block evolution of range*

For each stage an uneroded and an eroded block are shown. *Th*—Miocene Humboldt formation; *Pt*—pre-Tertiary rocks.

diately west of the fault, plus an unknown thickness of pre-Miocene rock eroded from the west slope of the mountain block. The displacements on the west boundary fault are not given by the difference in altitude between the crest of the range and the base of the Miocene deposits immediately west of the boundary fault because of westward tilting of the mountain block.

Table 1 gives minimum values for the vertical component of the dip-slip displacement on the boundary faults at a number of localities. The

TABLE 1.—*Minimum values for vertical component of dip-slip displacements*

Locations, from north to south	Vertical component of dip-slip displacements (feet)	Assumed thickness of Miocene deposits (feet)
East side of East Humboldt Mountains		
Ralph Creek	4800	1000
Leach Creek	5800	1000
Johnson Creek	4500	700
Polar Star mine	2200	500
East side of Ruby Mountains		
Robinson Creek	4200	No deposits known
Smithers Creek	4800	No deposits known
Wines Creek	4600	No deposits known
West side of Ruby-East Humboldt Range		
Boulder Creek	2000(?)	1000
Secret Creek	1000(?)	1000
Thorpe Creek	2000(?)	200

pre-Miocene surface is assumed to be one of low relief, an assumption supported by the fine-grained texture of the basal Miocene deposits. The thickness of the basin deposits assumed at each locality is shown in the table. The values are approximate, and some of them may be in error by a thousand feet or more. The displacements for the west side of the range are all given as questionable and represent the best estimates that can be made with the data at hand. The displacements for the east side of the range are given with considerably greater confidence, although the actual displacements may be much larger than the values given. The relative proportions between the displacements on the east and west sides are believed to be approximately correct, and the range has not only been uplifted as a horst but it has been tilted westward.

TERMINATING STRUCTURE OF RUBY-EAST HUMBOLDT RANGE

The terminating structure of the Basin Ranges has long been a problem. The two conceptions generally held are: (1) Termination by decreasing displacements on the boundary faults, so that in longitudinal

section the range looks like a broad anticline or arch, highest in the middle and lower toward both ends, and (2) termination by a cross fault or series of cross faults. The first view has been championed chiefly by Louderback (1904, p. 338; 1907-1908, p. 667) and the second by Schneider (1925, p. 47).

The terminating structure of the Ruby-East Humboldt Range has been studied only at the north end near Wells. A structure somewhat different from either of the two outlined has been found. The crest of the range maintains its usual height—10,000 to 11,000 feet—to within 3 or 4 miles of the north end of the mountains. A decreasing displacement on the boundary faults in the last 3 or 4 miles accounts for only part of the termination of the range. The map (Pl. 1) shows that the east boundary fault swings westward and the west boundary fault swings eastward at the north end of the range, so that the two faults approach each other, and the mountain block narrows progressively between them. At the northern tip the faults intersect, and the range is terminated. On the map, the west boundary fault is shown as crossing and terminating the east boundary fault; a low scarp along the line of the west boundary fault beyond the intersection of the two faults shows that the latest movement has been on the west boundary fault. At earlier periods, movements on the two faults may have been essentially contemporaneous.

PERIODS OF FAULT MOVEMENTS

Four periods or stages of Basin-Range faulting seem well established in the Ruby-East Humboldt region, and there is some evidence of a fifth and earlier stage. These are:

(1) Pre-Humboldt (early upper Miocene) faulting or warping creating the Miocene areas of deposition.

(2) Faulting during deposition of the upper Miocene Humboldt formation which is responsible for breccias and fanglomerates within that formation (later upper Miocene period).

(3) Post-Humboldt—pre-Pliocene (?) lava faulting. This period may extend from latest Miocene into the Pliocene.

(4) Post-Pliocene (?) lava faulting. This period may occupy any part of an interval starting sometime in the Pliocene and extending well into the Pleistocene. This has been the period of last major uplift of the range.

(5) Late Pleistocene to Recent faulting represented by piedmont scarps which are known to be younger than earliest Wisconsin and possibly younger than later Wisconsin.

The basin or basins in which the upper Miocene Humboldt formation accumulated may have originated either by faulting or warping. A period of early upper Miocene faulting to produce these basins is a possibility

which is as yet unproved. This would be the earliest period of Basin-Range faulting in this region.

Breccias and fanglomerates, interbedded in the Humboldt formation just south of Wells, have been derived from an actively growing fault scarp—the east face of the East Humboldt Mountains—a mile to the west. The relatively fine-grained beds above and below the coarse breccias and fanglomerates show that they represent a distinct break in the conditions of sedimentation, as described in more detail in another paper (Sharp, 1939). Comparison with similar deposits described by Longwell (1936, p. 1420-1429; 1937, p. 434-438) and Ferguson (1924, p. 377-378) shows that the theory of derivation from an actively growing fault scarp is entirely acceptable. These deposits indicate movement on the faults in upper Miocene time. This second period of faulting is separated from the first by a period of relatively quiet sedimentation. Several periods of faulting may have ensued during the deposition of the Humboldt formation, but evidence for only one is found.

The fault contacts between the Humboldt beds and the pre-Miocene rocks of the mountains give proof of at least one, and possibly several, periods of post-Humboldt faulting. In the Elko Range near Elko, and at the southern end of the East Humboldt Mountains, lava flows and breccias overlie unconformably the tilted and eroded Humboldt beds. A period of deformation and erosion ensued between the deposition of the Humboldt beds and the extrusion of the lavas. Buwalda (1914), Louderback (1904; 1924), and Ferguson (1924) among others have noted that the Miocene-Pliocene interval, and the early Pliocene, were marked by deformation and erosion in the western part of Nevada. Similar relations may have obtained in the northeastern part of Nevada. Therefore, the lavas are tentatively dated as Pliocene (?), and a period of deformation, presumably faulting, younger than the Humboldt and older than the Pliocene (?) lava is indicated. This may be called, for convenience, the Miocene to Pliocene period, for it may have occurred any time in the interval from late Miocene to sometime in the Pliocene.

The nature of the erosion surface upon which the Pliocene (?) lavas rest shows that the Ruby-East Humboldt region was reduced to an area of moderate or low relief prior to the extrusion of the lavas. The topography of the Ruby-East Humboldt Range is too sharp and too fresh, particularly the eastern face, to be attributed to the Miocene to Pliocene period of faulting, and a fourth period of faulting accounting for the last major uplift of the range is postulated. Unfortunately, the Pliocene (?) lavas are nowhere directly broken by the boundary faults. However, in the Elko Range, the Pliocene (?) lavas rest on top of the range in relations which indicate that they were extruded before the Elko

Range—a fault block—was uplifted to its present position. The available evidence indicates a period of faulting which may have occurred any time in an interval extending from some place in the Pliocene into the Pleistocene.

The Pliocene to Pleistocene period of faulting did not extend through the Pleistocene, for it was separated from the final period by an interval of stability during which a series of pediments and terraces were cut along the west flank of the range. Some of these surfaces are older than the Wisconsin glaciation, and they may be as old as middle Pleistocene, but they are definitely younger than the Pliocene to Pleistocene period of faulting.

The latest period of faulting is represented by piedmont scarps on both sides of the range. On the west side, these scarps cut drift which has been correlated with the Iowan substage of the Wisconsin glaciation[4] (Pl. 5, fig. 3). On Heenan Creek, south of Lamoille, outwash gravels of a later Wisconsin glaciation may or may not be cut by a piedmont scarp. The relations at the mouth of Heenan Creek could be explained by stream cutting, so no definite assertion can be made as to the date of the latest faulting in this area. It is certainly later than the Iowan substage of the Wisconsin and may possibly be post-Wisconsin. This period of faulting may be dated as late Pleistocene to Recent.

Faulting may have been continuous from the Miocene to the present, and the so-called periods outlined merely represent stages of accelerated faulting, though the evidence of extended erosion between stages suggests that at least some of them were separated by periods of stability. The late Pleistocene to Recent faulting at the west base of the range may show that it is being differentially uplifted as rapidly at present as at any time in its history.

Ferguson (1924) outlined four periods of Basin-Range faulting in the Hawthorne and Tonopah quadrangles of western Nevada, and detailed studies in other parts of the Great Basin may show a somewhat similar succession of periods.

Figure 8 is a series of diagrammatic cross sections which illustrate the evolution of the Ruby-East Humboldt Range as a Basin Range from the Miocene to the present. The pre-faulting surface is assumed to be a surface of low relief. The initial deformation is assumed to be faulting, though no definite proof of this is forthcoming, and it may have been warping. In each pair of sections, the assumption is made that all erosion takes place after and not during the faulting—an incorrect assumption, but one which does not materially alter the relations as

[4] The Pleistocene glacial succession outlined by Thwaites (1935, p. 72) is followed here.

shown in the sections. No faulting is postulated at the western base of the range until the second period. The exact time at which faulting was initiated at the west base is not known and may have been earlier or later than shown. Two small subsidiary faults are shown west of the principal fault at the west base. One such fault has been observed, and the highly deformed Miocene beds west of the range suggest that others are present.

The time of initiation of Basin-Range faulting has been determined variously as late Tertiary, Miocene, middle Miocene, late Miocene, Pliocene, early Pliocene, and late Pliocene by Schneider (1925), Gilluly (1928), Jenney (1935), Louderback (1924), Bryan (1936), Ferguson (1924), Knopf (1918), Eaton (1932), and others. From the foregoing discussion it is evident that Basin-Range faulting along the Ruby-East Humboldt Range was initiated in the upper Miocene. This falls within the limits set by the earlier studies mentioned.

The last major uplift of the Ruby-East Humboldt Range occurred in the interval extending from some time in the Pliocene to middle or upper Pleistocene. Eardley (1934, p. 396-397) has presented evidence which shows that the last major uplift of the southern Wasatch Mountains was late Tertiary or early Pleistocene. Eaton (1932, p. 39) has suggested that the block faulting of the Great Basin reached a maximum in the upper Pleistocene. Knopf (1918, p. 13, 88) states that the major uplift of the Sierra Nevada took place at the beginning of the Quaternary. Louderback (1924) has concluded that the period of maximum faulting in the western Great Basin was late Pliocene or early Pleistocene. R. J. Russell (1928, p. 495) concurs with this view.

In summary, the period of maximum block faulting in the Great Basin seems to have been latest Tertiary and Quaternary. The Pliocene to Pleistocene period of last major uplift of the Ruby-East Humboldt Range fits in well with this concept.

PIEDMONT SCARPS

GENERAL STATEMENT

The Great Basin as a whole, and Nevada in particular, might be called the type locality for piedmont scarps, which is the name given by Gilbert (1928, p. 34) to low scarps which break the piedmont slope and are related to dislocation. Longwell (1930, p. 1) proposed that fan scarp would be a better term. Not all piedmont scarps cut fans, and, since the term piedmont scarp seems so well intrenched in the literature, it will be used here.

Piedmont scarps may be located at the base of the principal scarps or some distance out on the piedmont slope. They have been described

from a great number of places in the Great Basin by Gilbert (1890; 1928), I. C. Russell (1882; 1885; 1886-1887; 1887), Baker (1912), Knopf (1918), Louderback (1904, 1907-1908), Lawson (1912), Lindgren (1911), Davis (1905), Hobbs (1910), Longwell (1930), R. J. Russell (1928), Taber

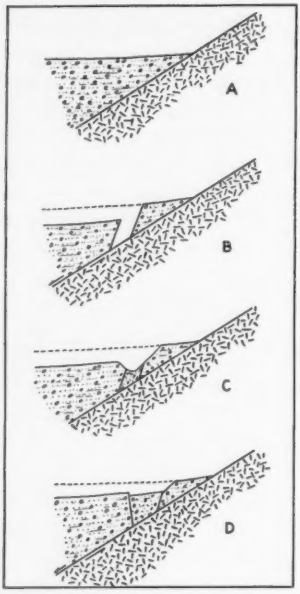

FIGURE 9.—*Piedmont scarps after Gilbert*

Illustrating how a piedmont scarp and the associated graben-like depression can be formed.

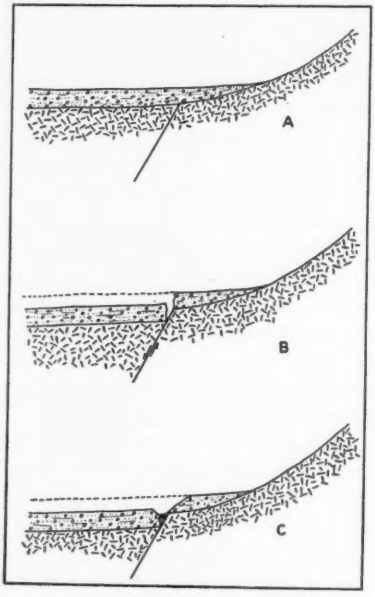

FIGURE 10.—*Piedmont scarps, modified*

Illustrating formation of a piedmont scarp and the associated graben-like depression, modified from Gilbert's scheme (*cf.* Fig. 9).

(1923), Page (1935), Gianella (1934), Eardley (1934), Anderson (1937), Gilluly (1928), Pack (1926), Jenney (1935), Schneider (1925), Reiche (1938), and others.

PIEDMONT SCARPS OF RUBY-EAST HUMBOLDT RANGE

Piedmont scarps are prominent along the west base of the Ruby-East Humboldt Range from the northern end southward approximately to Lee. Two piedmont scarps have also been mapped on the east side of the range.

At the mouth of Greys Creek near the north end of the East Humboldt Mountains on the west side is a piedmont scarp, 75 feet high, which cuts fan gravels (Pl. 5, fig. 1). The lower part of the scarp may be bedrock,

mantled by slumped gravel. About 75 feet in front of this scarp, parallel to it, is a low gravel scarp, 20 feet high, which faces mountainward. A low, graben-like depression separates the two scarps (Pl. 5, fig. 2). About 75 feet beyond this second scarp is another lower and more subdued mountainward-facing scarp. These scarps extend for several hundred feet parallel to the mountain front, and, since they are clearly not features of stream erosion, they may be attributed to recent movement on the principal boundary fault. Gilbert (1928, p. 36-37) proposed an explanation for these graben and subsidiary piedmont scarps, and his explanation has been accepted by Longwell (1930) and Page (1935). Taber (1923, p. 667) has criticized Gilbert's explanation, and some further objections are outlined here.

In Gilbert's figures the fan gravel or alluvium rests directly against a fault plane, the footwall of which is bedrock (Fig. 9, A). Whether the alluvium came into this position by an earlier fault movement or by deposition is not stated. When movement occurs on this fault, the hanging wall moves relatively upward carrying a part of the alluvial deposit with it (Fig. 9, B). The intervening trench is quickly filled, completely or in part by detritus from both sides (Fig. 9, C). Gilbert has also suggested that a slice or block of gravel may slump and spread during the fault movement and form a graben-like depression (Fig. 9, D). He further points out that some piedmont scarps are probably the direct continuation of faults in the underlying bedrock.

The explanation proposed by Gilbert needs modification on two points. In the figures copied from Gilbert the average dip of the fault plane is 33 degrees which is considerably lower than the attitude commonly observed in the Basin Ranges as pointed out by Gilluly (1928, p. 1113-1116), Pack (1926, p. 402-405), and Eardley (1934, p. 388). Furthermore, Gilbert assumes that the initial relations are those of alluvium resting against the fault face, which is rarely the case. Figure 10, A represents more truly the relations of alluvium, fault, and bedrock before faulting. The alluvium covers the fault and laps over onto the mountain block. Movement on the fault may produce a trench in the overlying gravel like that shown in Figure 10, B. Subsequent slumping and washing of gravel from both sides would fill the open trench in the same manner as outlined by Gilbert.

Some graben and piedmont scarps may be formed in the manner outlined, others probably represent directly breaks in the underlying bedrock. Complex distributive faulting would be expected in a wide fracture zone which forms the boundary fault of the mountain block, and within this zone slices or blocks may be relatively depressed, forming graben-like depressions in the overlying gravel.

Remnants of a piedmont scarp appear just south of Thorpe Creek and extend almost continuously for 9 miles south along the west base of the Ruby Mountains. This scarp cuts moraines correlated with the Iowan substage of the Wisconsin glaciation at the mouths of Lamoille, Seitz, and Heenan creeks (Fig. 11; Pl. 5, fig. 3). Outwash gravel associated with a

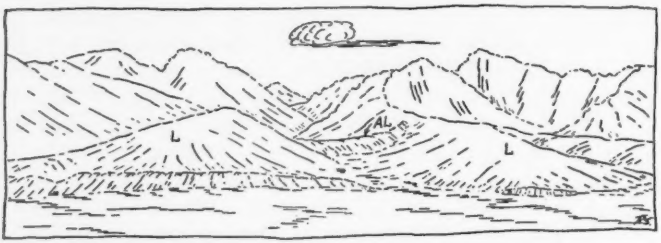

FIGURE 11.—*Sketch of piedmont scarp at Heenan Creek*

Scarp cuts outwash gravel of earliest Wisconsin glaciation. *L*—lateral moraines of the earliest Wisconsin; *AL*—terminal moraine of later Wisconsin. (Sketched from a photograph)

later substage of Wisconsin glaciation is cut by a 20-foot scarp at the mouth of Heenan Creek. This scarp is on a line with remnants of the piedmont scarp exposed at other places along the base of the range. However, it cannot be stated with certainty that the faulting is younger than the later Wisconsin gravel for the part of the scarp which cuts this gravel could have been carved by Heenan Creek.

On the east side of the Ruby Mountains, extending from Smithers Creek south to Battle Creek, a distance of 10½ miles, is a spectacular piedmont scarp a quarter of a mile to 1¼ miles east of the base of the range (Fig. 4). This scarp is well exposed near Lutts Creek and for the sake of brevity will be referred to as the Lutts scarp. The face of the Lutts scarp is composed of pre-Tertiary limestone intruded by pegmatitic and granitic rocks (Pl. 4, fig. 2). In places, overlying pediment gravel has slumped down over the face of the scarp and has covered the underlying bedrock. However, as far as can be seen, the scarp is nowhere composed of gravel, and in this way it differs from the piedmont scarps on the west side of the range. The Lutts scarp is highest between Lutts and Thompson creeks, ranging from 150 to 200 feet high. Just south of Lutts Creek the slope of the scarp is 30 degrees, but presumably the dip of the fault plane is much steeper.

A dissected pediment, cut on limestone and igneous rocks, lies between the Lutts scarp and the base of the mountains. No evidence of a fault has been found at the base of the mountain slope, although a diligent

search was made. This pediment and the Lutts scarp demonstrate a relation between the mountain face and the boundary fault often suspected but seldom proved (Fig. 12). It seems almost certain that the Lutts scarp

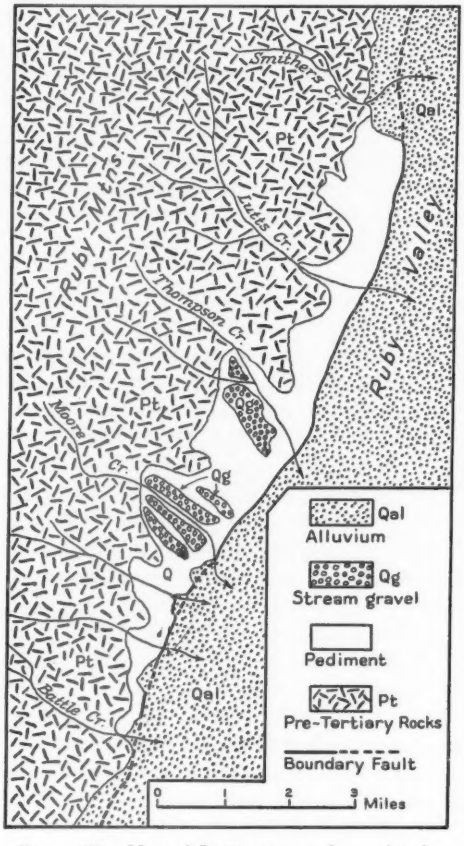

FIGURE 12.—*Map of Lutts scarp and associated pediment*

Showing relations of the boundary fault, pediment, and the base of the mountain slope on the east side of the Ruby Mountains.

represents a late displacement on the boundary fault and thus indicates the position of that fault. This means that the mountain front has been worn back to its present position, maintaining a steep slope and leaving

a pediment between it and the boundary fault along which the mountain block was first uplifted.

Terraces at approximately 50 and 100 feet above stream grade along Lutts Creek between the Lutts scarp and the mountain front, and a few terrace remnants on other streams indicate that the uplift of the Lutts scarp may have been in stages separated by short intervals of erosion.

The age of the Lutts scarp cannot be exactly determined. The pediment behind it is dissected to a depth of 200 feet by relatively large streams such as Lutts, Thompson, and Moore creeks. Small headward-working canyons have been cut in the scarp face and have extended back as much as half a mile (Fig. 4). The topographic features of the Lutts scarp suggest that it is not older than middle Pleistocene, though it does not appear to be as young as the late Pleistocene or Recent scarps on the west side of the range.

A scarp similar to, but less spectacular than, the Lutts scarp has been mapped farther south near Dawley Creek. This scarp extends for 1½ miles parallel to the base of the range and reaches a maximum height of 125 feet near the middle. The scarp face is composed of binary granite, and a pediment a quarter to a third of a mile wide has been cut on the granite between the scarp and the base of the mountain slope. This scarp is particularly notable for the great number of springs which emerge along its base. The statements concerning the age and nature of the Lutts scarp appear to apply to this scarp too.

Farther south, toward Harrison Pass, a number of low escarpments cut the piedmont slope near the base of the range. These escarpments might be mistaken for piedmont scarps, but study has shown them to be features of lake-shore erosion, associated with the lake which filled Ruby Valley in late glacial or postglacial time.

NATURE OF DIASTROPHIC FORCES

Recently, emphasis has been placed on the influence of compression in at least the western part of the Great Basin.

Smith (1927) studied the Steens Mountain-Pueblo Mountain uplift, a Basin Range in southeastern Oregon. He concluded that this range is bounded by thrust faults; however, Fuller and Waters (1929) argue for a typical horst and graben structure in the Steens Mountain region and seem to have disposed of the best of Smith's arguments.

Lawson (1936) recently suggested that the last major uplift of the Sierra Nevada has been at least indirectly due to thrusting. Mayo (1937) concluded that some of the uplifts of the southern Sierra Nevada may have been along steep upthrusts caused by compression. In neither of these studies have thrust faults actually been observed at the base of the range.

In the Ivanpah quadrangle, California-Nevada, along the western border of the Great Basin, Hewett (1928a; 1928b) found that Miocene basin deposits have been overridden by thrust blocks of older rocks. Baker (1911, p. 352) shows closed folds in the Miocene (?) Rosamond series of the Mojave Desert at the edge of the Basin-Range province, a fact which seems to have influenced his opinion that many of the ranges of the Great Basin are mountains of tangential compression (Baker, 1913). These areas are closely related to the Garlock fault, a branch of the San Andreas fault. Thrust faults are known along the San Andreas (Buwalda, 1935; Willis, 1938a, b). In addition, thrusting in the California Coast Ranges is extremely common. It seems entirely possible that the tectonic influences of the California Coast Range province may have overlapped the western part of the Great Basin and complicated the relations by Tertiary thrusting.

The thrust faults described by Hewett (1928a; 1928b, p. 10) are broken by later normal faults with displacements of a few hundred feet or less. Hewett does not state whether or not he considers the present ranges of the Ivanpah quadrangle to be related to similar but larger normal faults which are also younger than the mid-Tertiary thrusts.

In the Ruby-East Humboldt region there is no evidence of intensive late Tertiary or Quaternary compression. The beds of the upper Miocene Humboldt formation have been deformed by tilting, gentle warping, and local compression associated with the tilting and shifting of the mountain and basin blocks, not by intensive compression. All the late Tertiary or Quaternary faults observed in this region are normal faults. The relative displacements of the mountain and basin blocks appear to have occurred in a neutral or tensional state of the crust. The tilting of the blocks and the upward or downward movement of the mountain and basin blocks suggest that differential vertical stresses have played a part and possibly the major part.

Middle or late Tertiary thrusting in at least the western part of the Great Basin seems to be established beyond doubt, but the proof that the Basin Ranges are uplifted along thrust faults is yet to be furnished. All the evidence obtained in this study indicates that at least the Ruby-East Humboldt Range has been relatively uplifted by displacements on normal faults.

COMPOSITE NATURE OF SCARPS

Blackwelder (1928) has listed the criteria by which fault scarps and fault-line scarps may be distinguished, and his paper has stimulated interest in fault-line scarps. The geologic relations in the Ruby-East Humboldt Range are favorable to the formation of fault-line scarps. The basins are underlain by relatively soft, unconsolidated Miocene deposits,

and the upfaulted mountain block is composed of resistant pre-Miocene rocks. The relations are complicated by many factors, chief among which are the unknown amount of overlap of the Miocene basin deposits on the mountain block before post-Miocene uplift of the range and the effects of the several separated periods of uplift.

At the northern end of the East Humboldt Mountains beds of soft Miocene basin deposits, dipping steeply eastward, crop out immediately east of the scarp, and a pediment, now dissected, has been cut across these deformed basin deposits. This pediment terminates sharply against the base of the scarp of pre-Miocene rocks which forms the face of the mountains. If this pediment is older than the faulting which has produced the scarp against which it terminates, the scarp is a true fault scarp. If this is the case, remnants of the pediment or surfaces graded to it should be found higher up on the mountain block. Such remnants have not been found, and the suggestion is made that the pediment is younger than the scarp against which it terminates. If the pediment is younger than the scarp, an unknown thickness of Miocene basin deposits has been removed from the forelying block, and part of the scarp has been exposed to view by erosion and not directly by faulting. A conservative estimate would be that not over 1000 feet of the scarp, here about 4000 feet high, and perhaps less, have been exposed by erosion. That part of the scarp exposed solely by erosion is a true fault-line scarp. That part which owes its origin directly to fault movement is a true fault scarp. Cotton (1922, p. 171) has suggested the term composite scarp for scarps which are in part fault-line scarps and in part fault scarps. In Figure 13, A represents an uneroded fault block in which the upthrown block is composed largely of granite and other resistant rocks, and at least the upper part of the basin block is composed of soft sedimentary beds. B represents this same block modified by erosion. Some of the soft sedimentary beds have been removed from the downthrown block exposing part of the fault plane. The mountain face is a composite scarp made up in part of a fault scarp and in part of a fault-line scarp as shown in the sketch.

The east face of the East Humboldt Mountains is a composite scarp, at least at the north end and possibly for its entire length. The east face of the Ruby Mountains west of Ruby Valley has all appearances of a true fault scarp. It may be a composite scarp, but pertinent data are lacking.

The basins east of the range are hydrographically closed; nevertheless it seems likely that a considerable amount of material, chiefly soft Miocene deposits, has been removed from these basins. Some of this material may have gone westward into the Humboldt River, or, perhaps, some of it may have been moved northward by streams draining into the Salmon Falls, Bruneau, or Owyhee rivers, tributaries of the Snake. The drain-

age divides between these basins and the tributaries of the Snake River are only a few hundred feet high, and no great change need be postulated to account for disintegration of the drainage to its present state. Such a change could be attributed largely to tectonic activity and perhaps, in minor part, to climatic change. However, this is mere hypothesis, for

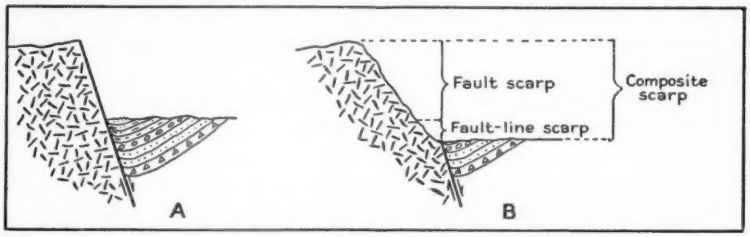

FIGURE 13.—*Sketch of a composite scarp*
A—the faulted and uneroded block; B—the eroded block.

in the Ruby-East Humboldt region there is no good evidence suggesting that such a drainage system ever existed.

The data on the western face of the range are more complete because of better exposure of the Miocene deposits in the basin to the west. Near the mouth of Thorpe Creek is a small eastward-tilted, forelying block on the back side of which the base of the Humboldt formation is exposed. It is known that the beds of the Humboldt overlap the pre-Miocene rocks of the mountain block, so that the Humboldt at Thorpe Creek was probably much thinner than its maximum known thickness of 5800 feet. However, only 200 feet of the Humboldt is exposed at the present time, and it seems entirely probable that its original minimum thickness at Thorpe Creek could not have been much less than 1000 or 2000 feet. A pediment truncates both Miocene and older rocks and indicates considerable post-faulting erosion. Some hundreds to perhaps as much as 1000 or 2000 feet of soft Tertiary basin deposits have been removed from the basin side of the scarp in this area. The total displacement on the west boundary fault at Thorpe Creek is about 2000 feet (?). Therefore, a large part and perhaps all of the scarp at Thorpe Creek may be a fault-line scarp. However, the original thickness of Miocene beds on Thorpe Creek is not known, and it seems best to consider this scarp as also of the composite type.

Remnants of pediments on the mountain block, 500 to 700 feet above the base of the range, have been observed on the west side of the Ruby Mountains. These pediment remnants have been correlated with more extensive and better-preserved pediments along other parts of the range

which extend unbroken from the basin deposits onto the mountain block. Detailed study shows that the dissection of the pediments and terraces on the west side of the range is due to rejuvenation of the Humboldt River drainage by some cause outside of this area and is not related to differential displacement of the mountain and basin blocks. The pediment remnants on the mountain block are in their present isolated position not because of uplift of the range but because of erosional removal of the soft Miocene beds in the basin. In these places, at least 500 to 700 feet of the scarp must be a fault-line scarp.

In summary, both the eastern and western faces of the Ruby-East Humboldt Range are composite fault scarps, hence not strictly true fault scarps or fault-line scarps. It is likely that this relationship is fairly general in the Great Basin and that other scarps are composite scarps, especially in those ranges having adjoining basins filled with soft Tertiary deposits. Horberg (1938, p. 64) has recently pointed out that the east face of the Teton Range—a fault block—outside of the Great Basin is probably a composite scarp.

Cotton (1922, p. 161) has suggested that the term rejuvenated scarp be applied to a scarp effected by recent displacements. Thus, the scarps on both sides of the Ruby-East Humboldt Range are, in places, rejuvenated scarps.

SUMMARY

The Ruby-East Humboldt Range consists of a complex of igneous, metamorphic, and sedimentary rocks. Four formations have tentatively been recognized in the metamorphic and sedimentary rocks. These are: Lower dolomite, 5000 feet thick; Quartzite formation, 4000 feet thick; Middle limestone, 4000 feet thick; and the Upper limestone, 5000 feet thick. These formations compose a continuous and conformable series, the two upper members of which are Carboniferous, and all of which is believed to be Paleozoic. These rocks have been folded extensively and intimately intruded by Late Jurassic or early Tertiary porphyritic granite, binary granite, and pegmatite. The basins adjoining the mountains contain deformed beds of the upper Miocene Humboldt formation, which is composed of continental deposits 5800 feet thick. Pliocene (?) lavas rest unconformably on the eroded edges of the deformed Miocene and older rocks.

The internal structure of the northern half of the range is a relatively simple series of broad, open folds, slightly asymmetric to the east. In the southern part of the range, structures are more complex and of diverse trends. A flat-lying thrust fault in Secret Pass is thought to be early Tertiary (?).

The internal structure is discordant with the trend and shape of the range. Folds within the mountain block are cut off obliquely or longitudinally by the faces of the range. This discordance is particularly clear on the east side. On the west side, pre-Tertiary structures may have influenced the trend of the boundary fault to some extent.

The range is a westward-tilted horst bounded on both sides by normal faults which dip 60 to 70 degrees basinward. Displacements on the east boundary fault are at least twice as great as on the west boundary fault. A minimum value for the vertical component of the greatest dip-slip displacement on the east side of the range is between 5500 and 6000 feet, and on the west side about 2000 (?) feet. Wherever observed, the boundary faults are normal. The northern termination of the range is due to intersection of the east and west boundary faults.

Four, and possibly five, periods of Basin-Range faulting have been recognized in this area as follows:

(1) Early upper Miocene (questionable);
(2) Later upper Miocene;
(3) A period in the interval from late Miocene to sometime in the Pliocene;
(4) A period in the interval from late Pliocene to middle or upper Pleistocene;
(5) Late Pleistocene to Recent.

The last major uplift of the range is thought to have occurred in the late Pliocene to middle or upper Pleistocene period.

Piedmont scarps which cut drift correlated with the Iowan substage of the Wisconsin glaciation are prominent features at the west base of the range. Older and higher piedmont scarps have been mapped on the east side of the range in two places.

No evidence of intensive late Tertiary or Quaternary compression and thrusting was found in the range or in the adjoining basins. Differential vertical forces, operating in a neutral or tensional state of the crust, seem to be indicated as the chief cause of the Basin-Range faulting in this area.

The scarps on both the east and west side are composite scarps, in major part to be attributed directly to fault movements, and in minor part related to erosional removal of soft beds from the downthrown block. In places the composite scarps are rejuvenated by late Pleistocene to Recent fault movements.

WORKS TO WHICH REFERENCE IS MADE

Anderson, G. H. (1937) *Granitization, albitization, and related phenomena in the northern Inyo Range of California,* Geol. Soc. Am., Bull., vol. 48, p. 1-74.

Baker, C. L. (1911) *Notes on the later Cenozoic history of the Mojave Desert region in southeastern California,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 6, no. 5, p. 333-383.

——— (1912) *Physiography and structure of the western El Paso Range and the Southern Sierra Nevada,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 7, no. 6, p. 117-142.

——— (1913) *The nature of the later deformation in certain ranges of the Great Basin,* Jour. Geo., vol. 21, p. 273-278.

Blackwelder, Eliot (1928) *The recognition of fault scarps,* Jour. Geol., vol. 36, no. 4, p. 289-311.

——— (1931) *Pleistocene glaciation in the Sierra Nevada and Basin Ranges,* Geol. Soc. Am., Bull., vol. 42, p. 865-922.

——— (1934) *Supplementary notes on Pleistocene glaciation in the Great Basin,* Washington Acad. Sci., Jour., vol. 24, p. 217-222.

——— (1935) *Summary of the pre-Cambrian rocks of Utah and Wyoming,* Utah Acad. Sci., Arts, and Letters, Pr., vol. 12, p. 153-157.

Bryan, Kirk, and McCann, F. T. (1936) *Successive pediments and terraces of the upper Rio Puerco in New Mexico,* Jour. Geol., vol. 44, no. 2, p. 145-172.

Buwalda, J. P. (1914) *Tertiary mammal beds of Stewart and Ione valleys in west-central Nevada,* Univ. Calif. Publ., Bull., Dept. Geol., vol. 8, p. 335-363.

——— (1935) *Investigation of overthrust faults by seismic methods,* Science, n. s., vol. 81, p. 384-386.

Callaghan, Eugene, and Gianella, V. P. (1935) *The earthquake of January 30, 1934, at Excelsior Mountains, Nevada,* Bull. Seismol. Soc. Am., vol. 25, no. 2, p. 161-168.

Cotton, C. A. (1922) *Geomorphology of New Zealand, part I, Systematic,* New Zealand Board Sci. Art, Manual no. 3.

Davis, W. M. (1903) *Mountain ranges of the Great Basin,* Mus. Comp. Zoology, Harvard College, Geol. ser., Bull., vol. 6, p. 129-177.

——— (1905) *Wasatch, Canyon, and House ranges, Utah,* Mus. Comp. Zoology, Harvard College, vol. 49, Geol. ser., Bull., vol. 8, p. 15-56.

——— (1913) *Nomenclature of surface forms on faulted structures,* Geol. Soc. Am., Bull., vol. 24, p. 187-216.

——— (1925) *The Basin Range problem,* Nat. Acad. Sci., Pr., vol. 11, p. 387-392.

——— (1927) *The rifts of southern California,* Am. Jour. Sci., 5th. ser., vol. 13, p. 57-72.

——— (1930) *The Peacock Range, Arizona,* Geol. Soc. Am., Bull., vol. 41, p. 293-313.

Dutton, C. E. (1880) *Geology of the High Plateau of Utah,* U. S. Geog. Geol. Survey Rocky Mountain Region, p. 47-48.

Eardley, A. J. (1934) *Structure and physiography of the southern Wasatch Mountains, Utah,* Mich. Acad. Sci., Arts, and Letters, vol. 19, p. 377-400.

Eaton, J. E. (1932) *Decline of the Great Basin, southwestern United States,* Bull. Am. Assoc. Petrol. Geol., vol. 16, no. 1, p. 1-49.

Ferguson, H. G., and Cathcart, S. H. (1924) *Major structural features of some western Nevada ranges,* Washington Acad. Sci., Jour., vol. 14, p. 376-379.

——— (1929) *The mining districts of Nevada,* Econ. Geol., vol. 24, p. 115-144.

Fuller, R. E., and Waters, A. C. (1929) *The nature and origin of the horst and graben structure of southern Oregon,* Jour. Geol., vol. 37, p. 205-238.

Gianella, V. P., and Callaghan, Eugene (1934) *The earthquake of December 20, 1932, at Cedar Mountain, Nevada, and its bearing on the genesis of Basin Range structure,* Jour. Geol., vol. 42, no. 1, p. 1-22.

Gilbert, G. K. (1874) *Preliminary geological report, expedition of 1872, U. S. Geog. Geol. Survey west of 100th meridian,* Prog. Rept., p. 50.

——— (1875) *Report on the geology of portions of Nevada, Utah, California, and Arizona, examined in the years 1871 and 1872,* U. S. Geog. Geol. Survey west of 100th meridian, vol. 3.

——— (1890) *Lake Bonneville,* U. S. Geol. Survey, Mon. 1.

——— (1928) *Studies of Basin Range structure,* U. S. Geol. Survey, Prof. Paper, 153.

Gilluly, James (1928) *Basin Range faulting along the Oquirrh Range, Utah,* Geol. Soc. Am., Bull., vol. 39, p. 1103-1130.

——— (1932) *Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah,* U. S. Geol. Survey, Prof. Paper 173.

Hewett, D. F. (1926) *Progress in the survey of the Ivanpah quadrangle, Nevada-California* (Abstract), Geol. Soc. Am., Bull., vol. 37, p. 164-165.

——— (1928a) *Two Tertiary epochs of thrust faulting in the Mojave Desert, California* (Abstract), Geol. Soc. Am., Bull., vol. 39, p. 178-179.

——— (1928b) *Late Tertiary thrust faults in the Mojave Desert, California,* Nat. Acad. Sci., Pr., vol. 14, p. 7-12.

——— (1931) *Geology and ore deposits of the Goodsprings quadrangle, Nevada,* U. S. Geol. Survey, Prof. Paper 162.

Hill, J. M. (1916) *Notes on some mining districts in eastern Nevada,* U. S. Geol. Survey, Bull. 648, p. 54-67.

Hobbs, W. H. (1910) *The earthquake of 1872 in Owens Valley, California,* Beitrag für Geophysik, vol. 10, p. 352-385.

Holmes, Arthur (1928) *The nomenclature of petrology,* p. 111.

Horberg, Leland (1938) *The structural geology and physiography of the Teton Pass area, Wyoming,* Augustana Library Publ., no. 16.

Hulin, C. D. (1930) *Subsequent faulting in the Great Basin* (Abstract), Pan Am. Geol., vol. 54, p. 151-152.

Jenney, C. P. (1935) *Geology of the central Humboldt Range, Nevada,* Univ. Nev., Bull., vol. 29, no. 6.

King, Clarence (1877, 1878) *Geological exploration of the Fortieth Parallel,* vol. 1, Systematic geology (1878); vol. 2, Descriptive geology (1877); Eng. Dept., U. S. Army, Prof. Paper, no. 18.

Knopf, Adolph, and Kirk, Edwin (1918) *A geological reconnaissance of the Inyo Range and the eastern slope of the southern Sierra Nevada, California,* U. S. Geol. Survey, Prof. Paper 110.

Lasky, S. G. (1936) *Geology and ore deposits of the Bayard area, Central Mining district, New Mexico,* U. S. Geol. Survey, Bull. 870.

Lawson, A. C. (1912) *The recent fault scarps at Genoa, Nevada,* Siesmol. Soc. Am., Bull., vol. 2, no. 3, p. 193-200.

——— (1936) *The Sierra Nevada in the light of isostasy,* Geol. Soc. Am., Bull., vol. 47, p. 1691-1712.

LeConte, Joseph (1889) *On the origin of normal faults and of the structure of the Basin region,* Am. Jour. Sci., 3d ser., vol. 38, p. 257-263.

Lindgren, Waldemar (1911) *The Tertiary gravels of the Sierra Nevada of California,* U. S. Geol. Survey, Prof. Paper 73.

Longwell, C. R. (1921) *Geology of the Muddy Mountains, Nevada, with a section to the Grand Wash Cliffs in western Arizona,* Am. Jour. Sci., 5th ser. vol. 1, p. 39-62.

——— (1928) *Geology of the Muddy Mountains, Nevada,* U. S. Geol. Survey, Bull. 798.

Longwell, C. R. (1930) *Faulted fans west of the Sheep Range, southern Nevada,* Am. Jour. Sci., 5th ser., vol. 20, p. 1-13.

——— (1936) *Geology of the Boulder reservoir floor, Arizona-Nevada,* Geol. Soc. Am., Bull., vol. 47, p. 1393-1476.

——— (1937) *Sedimentation in relation to faulting,* Geol. Soc. Am., Bull., vol. 48, p. 433-442.

Louderback, G. D. (1904) *Basin Range structure of the Humboldt region,* Geol. Soc. Am., Bull., vol. 15, p. 289-346.

——— (1907-1908) *Geological features of the Truckee region east of the Sierra Nevada,* Geol. Soc. Am., Bull., vol. 18, p. 662-669.

——— (1923) *Basin Range structure in the Great Basin,* Univ. Calif. Publ., Bull. Dept. Geol. Sci., vol. 14, p. 329-376.

——— (1924) *Period of scarp production in the Great Basin,* Univ. Calif. Publ., Bull. Dept. Geol. Sci., vol. 15, p. 1-44.

Mayo, E. B. (1938) *Structure of the southern Sierra Nevada, California* [Abstract], Geol. Soc. Am., Pr. 1937, p. 98.

Nolan, T. B. (1935) *The Gold Hill mining district, Utah,* U. S. Geol. Survey, Prof. Paper 177.

Pack, F. J. (1926) *New discoveries relating to the Wasatch fault,* Am. Jour. Sci., 5th ser., vol. 11, p. 397-410.

Page, B. M. (1935) *Basin Range faulting of 1915 in Pleasant Valley, Nevada,* Jour. Geol., vol. 43, no. 7, p. 690-707.

Reiche, Parry (1938) *Recent fault scarps, Organ Mountain District, New Mexico,* Am. Jour. Sci., 5th ser., vol. 36, p. 440-445.

Russell, I. C. (1882-1883) *A geological reconnaissance in southern Oregon,* U. S. Geol. Survey, 4th Ann. Rept.

——— (1885) *Geological history of Lake Lahontan, a Quaternary lake of northwestern Nevada,* U. S. Geol. Survey, Mon. 11.

——— (1886-1887) *Quaternary history of Mono Valley, California,* U. S. Geol. Survey, 8th Ann. Rept., p. 261-394.

——— (1887) *Notes on the faults of the Great Basin and of the eastern base of the Sierra Nevada,* Philos. Soc. Washington, Bull., vol. 9, p. 5-7.

Russell, R. J. (1928) *Basin Range structure and stratigraphy of the Warner Range, northeastern California,* Univ. Calif. Publ., Bull., Dept. Geol. Sci., vol. 17, p. 387-496.

Sharp, R. P. (1938) *Pleistocene glaciation in the Ruby-East Humboldt Range, northeastern Nevada,* Jour. Geomorph., vol. 1, no. 4, p. 296-323.

——— (1939) *The Miocene Humboldt formation in northeastern Nevada,* Jour. Geol., vol. 47, no. 2, p. 133-160.

Schneider, Hyrum (1925) *A discussion of certain geologic features of the Wasatch Mountains,* Jour. Geol., vol. 33, p. 28-48.

Simpson, J. H. (1876) *Report of explorations across the Great Basin of the Territory of Utah for a direct wagon route from Camp Floyd to Genoa in Carson Valley, in 1859,* Eng. Dept., U. S. Army.

Smith, W. D. (1927) *Contribution to the geology of southeastern Oregon (Steens and Pueblo Mountains),* Jour. Geol., vol. 35, p. 421-440.

Spurr, J. E. (1901) *Origin and structure of the Basin Ranges,* Geol. Soc. Am., Bull., vol. 12, p. 247-270.

——— (1903) *Descriptive geology of Nevada south of the fortieth parallel and adjacent portions of California,* U. S. Geol. Survey, Bull. 208.

Taber, Stephen (1923) *Some criteria used in recognizing active faults,* Geol. Soc. Am., Bull., vol. 34, p. 661-668.

Thwaites, F. T. (1935) *Outline of glacial geology,* Edwards Bros. Inc., Ann Arbor, Michigan.

Westgate, L. G., and Knopf, Adolph (1932) *Geology and ore deposits of the Pioche district,* U. S. Geol. Survey, Prof. Paper 171.

Wheeler, G. M. (1889) *Report upon United States geographical surveys west of the 100th meridian,* vol. 1.

Willis, Bailey (1938a) *San Andreas rift, California,* Jour. Geol., vol. 46, p. 793-827.

——— (1938b) *San Andreas rift in southwestern California,* Jour. Geol., vol. 46, p. 1017-1057.

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FIGURE 1. GABBRO AT CONTACT WITH TONALITE
Poikilitic crystals of hornblende (dark) enclose andesine. Without analyzer (x 25).

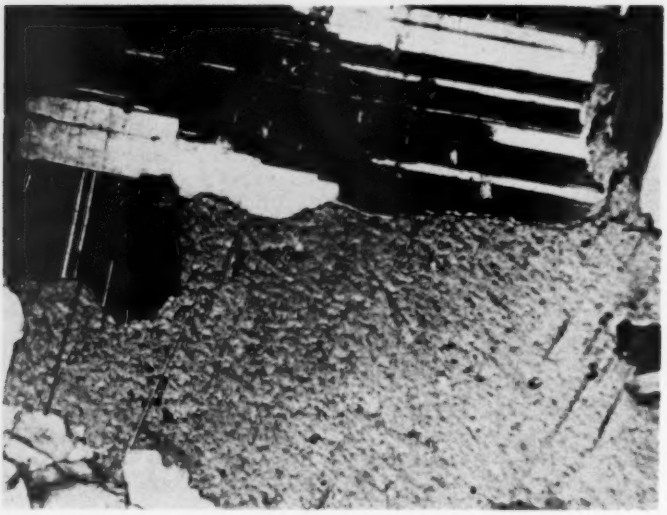

FIGURE 2. TONALITE
Tongue-like projections of biotite cut andesine. Crossed nicols (x 80).

PHOTOMICROGRAPHS OF GABBRO AND TONALITE

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STRUCTURAL PETROLOGY OF THE VAL VERDE TONALITE, SOUTHERN CALIFORNIA

BY E. F. OSBORN

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ABSTRACT

Micrometric analyses of the Val Verde tonalite were made on samples taken along a line extending from the border of the intrusive to a point 5 miles out in the intrusive. An attempt was made to evaluate sampling errors and to determine variation in mineralogic composition and radioactivity. The border of the intrusive was found to be more sodic and higher in radioactivity than the central part, and this is thought to be due to assimilation of quartz-biotite schist.

Inclusions in the tonalite are believed to be xenoliths of gabbro and quartz-biotite schist.

Petrofabric analyses of the schist country rock and the tonalite indicate that: (1) the schist was deformed primarily by rotation of mineral grains about a horizontal axis; (2) during intrusion of the tonalite a gneissoid structure developed in it parallel to the contact with the country rock with a linear element lying in the plane of foliation and paralleling the dip; (3) later stresses directed parallel to the strike of the foliation caused rotation of mineral grains in the tonalite about an axis parallel to the dip. The orientation of the minerals in the tonalite is thus believed to be due to a combination of magmatic flow and post-magmatic deformation.

INTRODUCTION

Two phases of the problem of batholiths and their origin are of special interest. The first concerns variation in composition of intrusive bodies, and the second deals with the origin and interpretation of primary flow structure. A unique opportunity to investigate these problems as applied to one intrusive was furnished by the construction of the Val Verde tunnel, one of a series composing the Colorado River Aqueduct which will carry water from Parker Dam on the Colorado River to Los Angeles and vicinity.

For slightly over 5 miles this tunnel passes through an apparently uniform tonalite intrusive at nearly right angles to the strike of the flow structure in the mass, and crosses the contact between the tonalite and quartz-biotite schist. The tonalite was examined along the line of the

tunnel for variations in mineralogic composition and radioactivity; petrofabric data were obtained from oriented specimens in an attempt to determine the origin of the gneissoid structure.

ACKNOWLEDGMENTS

The study was made possible by the friendly cooperation of the Metropolitan Water District of Southern California and of Mr. R. W. Remp, general manager for the Dravo Contracting Company. The problem was suggested by Professor Ian Campbell, who gave helpful advice throughout the investigation. Dr. J. R. Schultz, Mr. Cooper Hyde, and Mr. Ygnacio Bonillas, III, assisted the writer during part of the tunnel mapping. Dr. H. W. Fairbairn, Dr. James Gilluly and Dr. Earl Ingerson offered valuable suggestions on petrofabric procedure and interpretation. The radioactivity measurements were made by Dr. R. A. Clarke and Mr. F. H. Wright of the California Institute of Technology. The writer is indebted to Professor Ian Campbell, Dr. James Gilluly, Professor J. T. Stark, Professor C. H. Behre, Jr., Professor Adolph Knopf, and Dr. Earl Ingerson for their criticisms of the manuscript.

EARLIER PUBLICATIONS

The general geology and physiography of the Perris Block, of which the Val Verde district is a small part, are discussed in papers by Dudley (1935 and 1936). The heavy accessory minerals of the Val Verde tonalite have been described by Wilson (1937). In some of the early reports of the California State Mining Bureau (Fairbanks, 1892; Goodyear, 1888), and in a paper by Waring (1919), very general descriptions of the geology of this region are given. The surface along the line of the Val Verde tunnel was mapped by Ransome (1932) and Buwalda for the Metropolitan Water District of Southern California. Professor Larsen and students of Harvard University are engaged in field work in surrounding districts. Some of the results of their studies are referred to below.

LOCATION OF DISTRICT

The railroad station of Val Verde, from which the tunnel is named, is located in Riverside County, California, 12 miles south of the town of Riverside. Beginning about a mile south of Val Verde, the tunnel extends for 7 miles N.88°W. A strip of country extending about 3 miles on each side of the tunnel was mapped. This area of approximately 45 square miles occurs within the Riverside quadrangle of the Topographic Atlas of the United States. The location of the Val Verde district with respect to the principal structural features of southern California is shown in Figure 1.

GENERAL GEOLOGIC SETTING

GENERAL STATEMENT

The Val Verde district is part of a conspicuous topographic unit in southern California, named by English (1926, p. 54) the Perris Fault

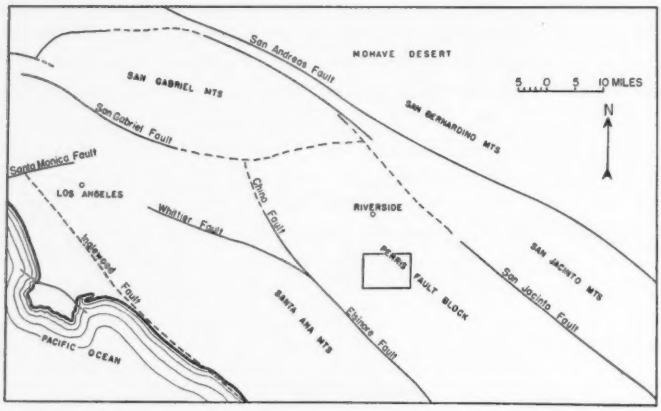

FIGURE 1.—*Map showing location of Val Verde district*
(From map prepared by W. A. English, 1926, U. S. Geol. Survey, Bull. 768)

Block. This unit is approximately 20 miles wide, bounded on the west by the Elsinore trough and on the east by the San Jacinto fault zone; it is of indefinite length, extending northward to the base of the San Gabriel Mountains and southward for some tens of miles. The region is one of relatively low relief, having an average elevation of about 1700 feet above sea level. The dominant rocks of the Val Verde district are medium-grained igneous rocks which have been intruded into a Triassic metamorphic series.

METAMORPHIC ROCKS

The oldest rocks in the Val Verde district are metamorphosed sediments, named by Dudley (1935, p. 403) the Elsinore metamorphic series. They occur as disconnected strips, elongated parallel to the strike of the beds. These remnants of what may have been formerly a continuous blanket of sediments range in length from a few hundred feet to several miles, and in width from a few tens of feet to about a mile. Within the Val Verde district, 12 such bodies of schist were mapped (Fig. 2).

Quartz-biotite schist is the dominant metamorphic rock and this grades into quartzite in a few places. The schist contains variable proportions of quartz, biotite, muscovite, hornblende, oligoclase, and orthoclase. Near

acidic igneous rocks, injection or replacement gneiss has been formed with very irregular and indefinite contacts. Small bodies of granitized schist occur in the intrusive, and apophyses of igneous rock protrude through the schist. Foliation in the schist is parallel to the bedding.

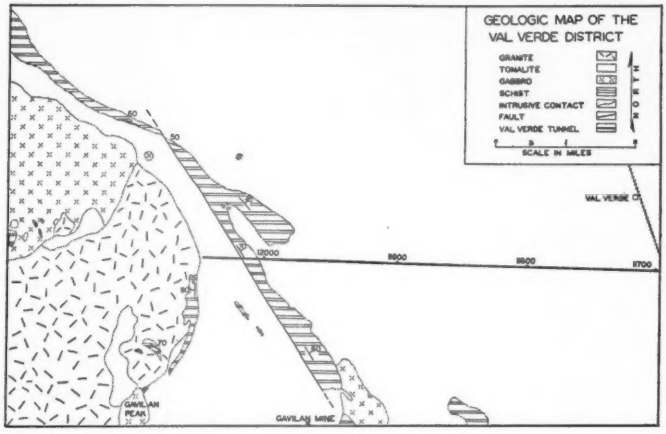

FIGURE 2.—*Geologic map of the Val Verde district*

Figures along line of tunnel refer to distance in hundreds of feet from Parker Dam on the Colorado River.

Based on lithologic correlation with the Triassic metamorphic series in the Santa Ana Mountains, 12 miles to the southwest, the age of the schist is believed to be Triassic.

IGNEOUS ROCKS

General statement.—In probable order of decreasing age the igneous rocks of the Val Verde district are gabbro, tonalite, granite, and basic and acidic dikes. These form a series intruded apparently during the same period of igneous activity.

Gabbro.—The igneous rock thought to be the oldest in the district is the Gavilan Peak gabbro, which crops out at five localities in the western part of the district. The rock is dark, massive, medium- to coarse-grained, and hypidiomorphic. Bytownite (An_{85-90}) constitutes 60 to 70 per cent of the rock, and augite, olivine, and hornblende, in variable proportions, make up the other 30 to 40 per cent. At some places near the contact with tonalite, olivine and augite are absent, their place being taken by hornblende and biotite which commonly enclose andesine poi-

kilitically (Pl. 1, fig. 1). The gabbro clearly intrudes the schist in the extreme western part of the district.

Tonalite.—The Val Verde tonalite, named by Ransome (1932), crops out over the entire eastern and central parts of the district, and occurs in the western part as isolated masses. On the basis of mapping by Dudley (1935),[1] tonalite is believed to underlie an area of at least 150 square miles on the Perris Block.

The tonalite is light-gray, medium-grained, and possesses a gneissoid structure which parallels in general the strike and dip of the schist. From the eastern end of the tunnel to the tonalite-schist contact, 5 miles to the west, it varies only slightly in composition and general appearance. The percentages of the minerals for an average sample of the rock are: andesine 57, quartz 26, biotite 12, hornblende 4, and orthoclase 1 per cent. Andesine and hornblende are subhedral, and biotite and quartz are anhedral. Most of the orthoclase is anhedral and closely associated with biotite. Commonly, however, small amounts of orthoclase occur as tiny rectangular blocks in the andesine aligned parallel to the cleavage directions of the host mineral. Biotite, quartz, and orthoclase are later than andesine and hornblende (Pl. 1, fig. 2). The accessory minerals, occurring in small amounts, are apatite, zircon, sphene, magnetite, pyrite, and rarely tourmaline. Wilson (1937, p. 124) reported also monazite and anatase. The average diameter of the grains is 1.0 mm., but a small percentage of the grains attain a maximum length of 3 to 5 mm.

The texture of the tonalite is dominantly igneous but the rock possesses two characteristics which indicate some deformation and recrystallization: (1) Contacts between mineral grains are commonly sutured, and, although cataclastic structures are not generally evident, andesine grains are locally bent or broken (Pl. 2, fig. 1). (2) The quartz not only commonly shows undulatory extinction but some grains are divided into several segments having slightly different extinction directions and sharp boundaries (Pl. 2, fig. 2). The "segmented" quartz grains may represent quartz which has not entirely recrystallized following deformation.

The tonalite intrudes the schist, but the contacts are gradational and irregular, suggesting that assimilation may have been important during emplacement of the tonalite. The tonalite grades into the schist through first a tonalitic gneiss containing bands high in quartz and biotite, and second a gneiss resembling the quartz-biotite schist but containing bands high in feldspar. In this transitional zone the foliation of the tonalite is parallel to the bedding planes of the schist.

[1] The Val Verde tonalite is the same as the Perris quartz diorite of Dudley.

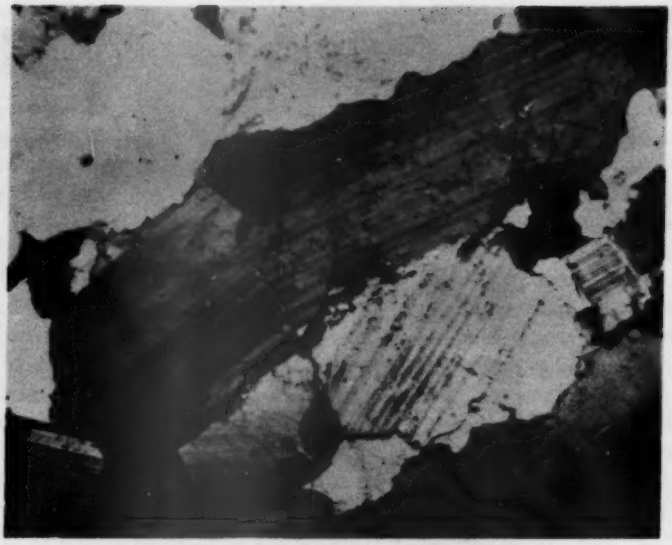

FIGURE 1. DISRUPTED ANDESINE CRYSTAL
Crossed nicols (x 50).

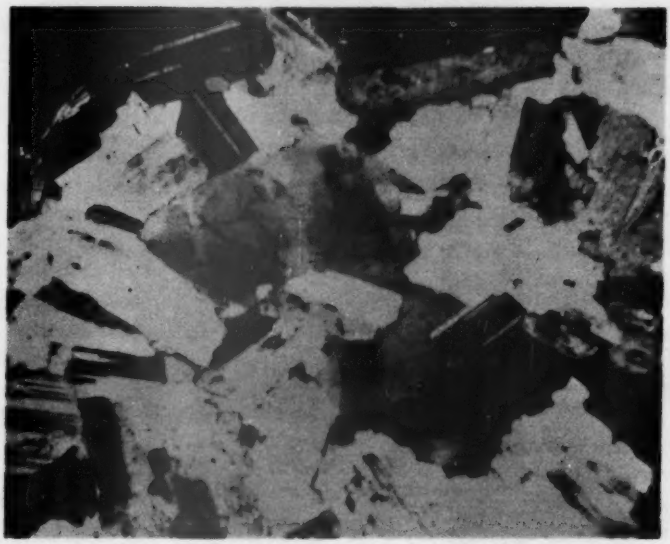

FIGURE 2. TWO ANHEDRAL GRANULATED QUARTZ GRAINS
Crossed nicols (x 35).

PHOTOMICROGRAPHS OF TONALITE

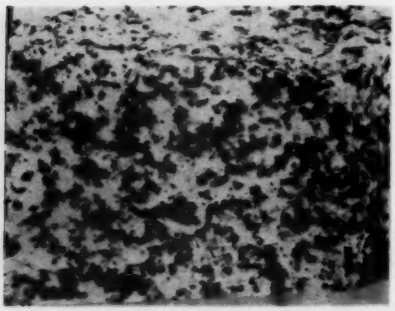

FIGURE 1. TONALITE
ac face. *a* is horizontal, *c* is vertical.

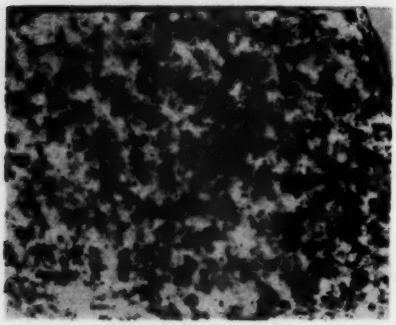

FIGURE 2. TONALITE
ab face. *a* is horizontal, *b* is vertical.

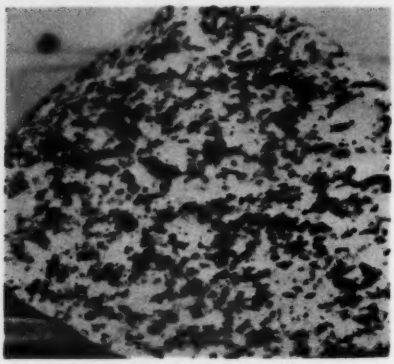

FIGURE 3. TONALITE
bc face. *b* is horizontal, *c* is vertical.

HAND SPECIMEN ORIENTED WITH
RESPECT TO AXES OF REFERENCE
Natural Size

In the few places in which contact relations with gabbro are unequivocal the tonalite intrudes the gabbro, but in some places the boundaries appear gradational and age relations are not clear.

Inclusions in tonalite.—Discoidal masses of dark, fine-grained rock, ranging in maximum dimensions from less than an inch up to several tens of feet, occur as inclusions in the tonalite, and are arranged parallel to the foliation. These dark bodies are absent over some areas as large as a few hundred feet in diameter, and in others constitute 60 per cent of the rock. Elongation of the inclusions within the plane of gneissosity is not evident, and the foliation of the tonalite is not noticeably curved around the inclusions but rather appears to pass through them. Boundaries between most of the inclusions and the tonalite appear sharp to the unaided eye and in thin section can be seen to be gradational over a distance of not more than 2 or 3 millimeters, but a small percentage of the inclusions blend into the tonalite to the extent that no definite boundary is evident. The inclusions are much more variable than the host rock, but as an average contain about 50 per cent plagioclase, ranging in composition from bytownite to oligoclase, 10 per cent quartz, and 40 per cent hornblende and biotite, hornblende commonly being more abundant than biotite. Orthoclase is present in only one inclusion examined. Apatite, zircon, magnetite, and rarely sphene are the accessory minerals. The average diameter of the grains is 0.25 mm., the size varying with the inclusion; but within a single inclusion the grain size is uniform. Grains are anhedral, contacts are notably sutured, and an order of crystallization is not apparent.

Granite.—Granite [2] which crops out in the western part of the district and also occurs as dikes in the tonalite, is light-brown, medium-grained, and possesses a gneissoid structure parallel to that in the tonalite. Dark, fine-grained inclusions are sporadically present, but are not generally as abundant as they are in the tonalite.

Determination of the composition of the granite is based on four specimens believed in the field to be typical of the particular locality from which they were taken. Quartz composes about 30 per cent of the rock, biotite about 5 per cent, oligoclase 5 to 20 per cent, and orthoclase and microcline 45 to 60 per cent.

At some localities the granite clearly intrudes the tonalite, but along the main tonalite-granite contact no definite boundary between the two rock types could be found.

[2] This granite is the eastern part of the intrusive named by Dudley (1935) the Cajalco quartz monzonite.

Dikes.—Numerous granite, aplite, and pegmatite dikes, and a few quartz monzonite, diorite, and quartz diorite dikes intrude the other rock types in the district. Most of these late intrusives are planar bodies accordant with the structure of the country rock. Some are plug-like masses roughly circular in plan.

Age and correlation.—All the igneous rocks intrude the Triassic metamorphic series and are earlier than the Eocene Alberhill clays (Dudley, 1935, p. 505). In the Santa Ana Mountains, which border the Perris Block on the west, igneous rocks similar to those in the Val Verde district are overlain unconformably by the Trabuco formation which is either late Lower, or early Upper, Cretaceous (Popenoe, W. P., personal communication, 1937). It seems probable, therefore, that the igneous rocks were emplaced during the Jurassic period, or at least during the Mesozoic era, and are comparable in age to intrusives in the Sierra Nevada.

STRUCTURAL GEOLOGY

The position of the Val Verde district with respect to the main structural features of southern California is shown in Figure 1. The district is located between two major strike-slip faults, the San Jacinto on the east and the Elsinore on the west. These faults strike approximately N.30°W. and are vertical.

The structural trend within the Val Verde district is outlined by the narrow body of schist which crosses it from northwest to southeast. In the southern and central parts of the district the schist strikes N. 20° to 30° W. and dips steeply to the northeast. In the northern part of the district the schist swings westward, assuming a N. 60° to 65° W. strike and a slightly lower angle of dip. The foliation in the tonalite and granite, the inclusions, and most of the dikes follow this same trend. The foliation in the tonalite, however, decreases in dip eastward from the contact with the schist and is essentially horizontal at the eastern end of the tunnel.

Numerous faults and joints cutting the tonalite occur in the tunnel. Most of these structures strike northwest and dip steeply. No constant relation between faults, joints, flow structure, and dikes could be established. If the fractures connected with the intrusion of the tonalite are present, they cannot be differentiated from later structures. The largest fault is one located at the western contact of the schist and tonalite. It is represented by a vertical zone of crushed and sheared rock and gouge 10 to 15 feet wide, striking N. 30° W. This fault (Fig. 2) has been traced on the surface for a distance of 4 miles, and the body of schist which it transgresses is offset approximately 500 feet, the west side having been shifted northwestward with respect to the east side.

Drag folds with an amplitude of a few inches occur sparingly in the schist. The axes of the folds are approximately horizontal. Vertical joints commonly cross the folds at right angles to the strike of the axes. Near the contact with the tonalite another type of fold occurs in the schist, the beds being commonly contorted and crinkled about an axis which parallels the dip of the schist.

MICROMETRIC AND RADIOACTIVITY MEASUREMENTS

GENERAL STATEMENT

An important part of the investigation was concerned with variation in mineralogic composition and radioactivity of the tonalite. The problem was of particular interest because of the unique opportunity to study fresh samples of a rock collected at regular intervals over a distance of 5 miles across a single intrusive. Study of the heavy minerals was made by R. W. Wilson of the geology department, and of the radioactivity by R. A. Clarke and F. H. Wright of the physics department of the California Institute of Technology.

The writer's contribution to the problem is divided into four sections: (1) examination of errors involved in sampling the rock; (2) determination of variation in mineralogic composition of the tonalite along the line of the tunnel; (3) correlation of these data with those obtained from heavy mineral and radioactivity studies; and (4) application of micrometric and radioactivity data to the problem of origin of the inclusions.

ERRORS IN SAMPLING VAL VERDE TONALITE

In a recent paper by Larsen and Miller (1935) errors involved in determining the mineralogic composition of a rock by a Rosiwal analysis of a thin section are discussed. Before examining the Val Verde tonalite for mineralogic variation, an attempt to evaluate sampling error was made by subjecting the rock to the various tests used by Larsen and Miller. Rosiwal analyses were made by using a Leitz six-spindle integrating stage, spacing the traverses at intervals of 1 millimeter. Aside from its bearing on the specific problem, the following description is presented in some detail because of the interest which it may hold for the student of similar problems.

In order to test the accuracy of the measuring method, two measurements were made on each of two thin sections of tonalite. On each thin section one measurement was made parallel to the trace of the plane of foliation, and the other measurement was made normal to this direction. The deviation from the average value of the two measurements

on a thin section was not greater than one per cent [3] for any constituent.[4]

To gain some idea as to the size of thin section necessary to give accurate results two tests were used. First, ten large thin sections were divided in half and the upper and lower halves measured separately, a method used by Larsen and Miller (1935, p. 264). The length of traverse

TABLE 1.—*Measurements on separate halves of two thin sections*

SPECIMEN 11982	*Lower*	*Upper*	*Weighted Average*	*Max. Deviation from the mean*
Length of traverse	272.6 mm.	231.4 mm.		
Plagioclase	67%	60%	64%	4%
Quartz	23	21	22	1
Orthoclase	1	1	1	0
Biotite	9	18	13	5
SPECIMEN 12024H				
Length of traverse	309.3 mm.	322.4 mm.		
Plagioclase	40%	39%	39%	1%
Quartz	27	28	28	1
Orthoclase	15	19	17	2
Biotite	9	9	9	0
Hornblende	9	5	7	2

across half a section averaged about 250 mm., or in other words, the area was about 250 sq. mm. The maximum deviation of any single constituent in one half from the average of the two halves was 5 per cent. The average deviation from the mean in per cent for the ten sections was, for plagioclase 2.1, for quartz 1.1, for orthoclase 1.4, for biotite 1.7, and for hornblende 0.8. In Table 1 are given the values for two of the thin sections. The first is the section possessing the least correspondence between the two halves, and the other is one showing about average correspondence.

Inasmuch as sections averaging 250 sq. mm. introduce a possible error of 4 to 5 per cent for an individual constituent, sections smaller than this were not considered reliable as samples of the rock.

Second, two thin sections larger than 300 sq. mm. were made from each of three hand specimens. The chips from which the sections were made were taken approximately two inches apart. Table 2 lists the results obtained from measurements of these thin sections.

The correlation shown in Table 2 is interesting in that two large thin sections of the same hand specimen agree more closely than two halves of the same large section (Table 1). The maximum deviation in a single thin section from the mean of the two sections of the same specimen

[3] In this and succeeding discussion, per cent are used as units when referring to deviation from a mean value. Thus, a deviation of 1 per cent from a mean value of 14 per cent indicates a possible fluctuation from a maximum of 15 to a minimum of 13 per cent.

[4] The data on these measurements are not given here inasmuch as they serve only as corroboration of other published investigations on the accuracy of the Rosiwal method. The data are contained, however, in Ph.D. Thesis of E. F. Osborn, California Institute of Technology, 1937.

is one per cent for plagioclase, quartz, orthoclase, and biotite, and 1.5 per cent for hornblende. This error is about the same as that of the measuring method itself; it is therefore concluded that one thin section larger than 300 sq. mm. is sufficient as a sample of a hand specimen

TABLE 2.—*Measurements of two thin sections from each of three hand specimens*

Specimen	*Ia*	*Ib*	*IIa*	*IIb*	*IIIa*	*IIIb*
Length of traverse	412.1 mm.	504.0	317.8	611.3	562.5	598.7
Plagioclase	63%	63%	52%	51%	42%	44%
Quartz	23	22	30	29	27	28
Orthoclase	1	1	1	1	14	14
Biotite	13	14	13	12	9	8
Hornblende	0	0	4	7	8	6

TABLE 3.—*Composition of eight samples of tonalite taken from area six feet in diameter*

COMPOSITION

Sample No.	*Length of traverse*	*Plagioclase*	*Quartz*	*Orthoclase*	*Biotite*	*Hornblende*
A	447.5 mm.	42%	27%	15%	12%	4%
B	519.6	44	28	13	9	6
C	562.5	42	27	14	9	8
D	460.3	44	29	11	11	5
E	354.5	34	25	27	8	6
F	598.7	44	28	14	8	6
G	503.2	43	29	17	8	3
H	631.7	40	27	17	9	7
Mean Value	.....	41.6	27.5	16.0	9.2	5.6

DEVIATION FROM THE MEAN VALUE

Sample No.	*Plagioclase*	*Quartz*	*Orthoclase*	*Biotite*	*Hornblende*
A	0.4%	0.5%	1.0%	2.8%	1.6%
B	2.4	0.5	3.0	0.2	0.4
C	0.4	0.5	2.0	0.2	2.4
D	2.4	1.5	5.0	1.8	0.6
E	7.6	2.5	11.0	1.2	0.4
F	2.4	0.5	2.0	1.2	0.4
G	1.4	1.5	1.0	1.2	2.6
H	1.6	0.5	1.0	0.2	1.4
Av. Dev. from Mean	2.3	1.0	3.2	1.1	1.2

of the tonalite. Most of the thin sections used in determining variation in composition of the tonalite were larger than 300 sq. mm. and none was smaller than 250 sq. mm.

The final test made was on the accuracy of a hand specimen as a sample of a larger body of rock. At station 12024 in the Val Verde tunnel (Fig. 2) eight specimens of what appeared to be typical tonalite were taken from an area six feet in diameter. Later study revealed that the rock was not tonalite, but that it varied in composition from granodiorite to quartz monzonite. Table 3 lists the composition of the samples as determined from one large thin section of each and gives also the

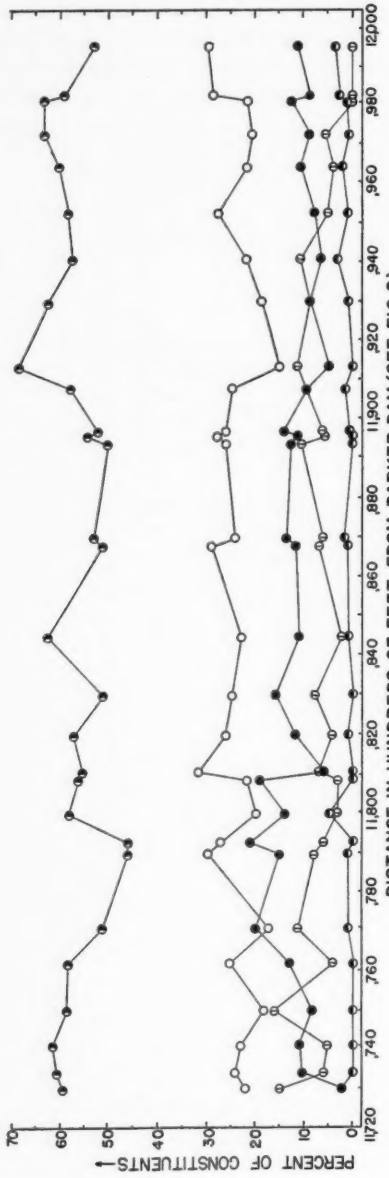

FIGURE 3.—*Graph of variation in mineral composition of tonalite*

average deviation from the mean for each constituent. The samples are very close in composition except for number "E", which has a rather different plagioclase-orthoclase ratio from the others. The mineral with the largest average deviation from the mean is orthoclase with a value of 3.2 per cent. If the tonalite along the tunnel is as uniform as the rock at this station, then variations in composition from point to point greater than about 3 per cent for a single constituent are probably outside the limits of error of the sampling method.

VARIATION IN MINERALOGIC COMPOSITION OF TONALITE

Twenty-nine sections have been measured to find the variation in composition of the tonalite along the line of the tunnel, and the results are shown in Figure 3. No mineral tends to increase or decrease regularly from east to west through the tunnel, although the tonalite near the west end contains in general slightly more orthoclase.

The variation in composition of plagioclase, however, is significant. As shown on the upper graph of igure 4, the plagioclase becomes more sodic toward the western end of the tunnel. Thus, the generalization may be made that the tonalite is more acidic toward the western end of the tunnel or toward the border of the intrusive. Wilson (1937, p. 129) came to the same conclusion from a study of the heavy accessory minerals. He found that toward the contact zircon increased in amount, apatite decreased, and there was an introduction of tourmaline and anatase.

The more acidic border of the tonalite is believed to be due to assimilation of quartz-biotite schist. Incorporation of this material into the tonalite magma would tend to make it more acidic, and contact relations indicate that assimilation may have been important. Radioactivity measurements, which are discussed in the next section, show that radioactivity of the tonalite increases toward the contact with the schist. This is suggestive of assimilation, since the schist as observed away from the contact averages higher in radioactivity than the tonalite.

CORRELATION OF MINERALOGIC DATA WITH RADIOACTIVITY DETERMINATIONS

Radioactivity determinations.[5]—Radioactivity measurements[6] were made on the tonalite from 35 different stations along the tunnel and on specimens of granodiorite, granite, schist, gabbro, inclusions, and dikes.

Results of measurements of the tonalite are plotted on the lower graph in Figure 4. Although the points are rather scattered, a curve has been drawn which correlates very well with that giving the albite content of

[5] In the following discussion the values listed for radioactivity are proportional to the number of alpha particles emitted per unit time per unit weight of sample. As far as is now known, the only elements involved are members of radium, thorium, and actinium series.

[6] A paper by Clarke and Wright is in preparation giving a description of apparatus and method and giving more detailed results than those listed herein.

the plagioclase along the tunnel. There may be some doubt as to the existence of the wave in the curves between stations 11760 and 11850, but the rise in both curves toward the west, or toward the contact with the schist, is real.

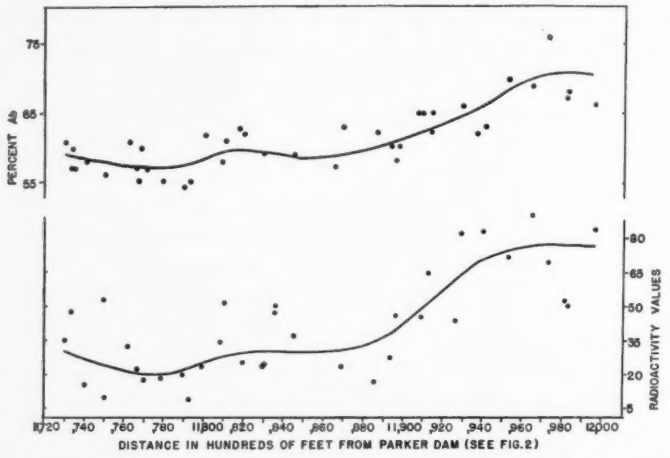

FIGURE 4.—*Graphs of albite content of plagioclase and radioactivity of tonalite*

A summary of the data concerning correlation of radioactivity with rock type is given in Table 4. These data show a direct relation between radioactivity and acidity of the rock.

Earlier studies of radioactivity of rocks.—Lord Rayleigh (Strutt, 1905) was the first to point out the direct relation between radioactivity and acidity of rocks. The data presented herein furnish additional proof of the correctness of the generalization. Since 1905 a series of investigations has dealt with the relation of radioactivity and chemical and mineralogic constitution of rocks. As examples, Waters (1909 and 1910) found that in the Cornish granite the radium was concentrated in anatase or rutile, in the Dalbeattie granite in allanite, and in the Mourne granite chiefly in zircon and some titaniferous mineral. Fletcher (1911), in studying the Leinster granite, found that the radioactivity of the rock was due mostly to micas containing inclusions of zircon. Hirschi (1927) in investigating the intrusive rocks of the Aar massif found a direct relation between the potassium content of the rocks and the radioactive elements of the uranium-thorium series. Piggot and Merwin (1932) examined two granites with a high content of radioactive substances, one

from Stone Mountain, Georgia, and the other from North Jay, Maine. As a result of their work, they concluded that in mica-bearing granites the radium is associated more with the micas than with other constituents of the rock. Evans and Williams (1935) concluded from a study of the

TABLE 4.—*Correlation of radioactivity and rock type*

Number of Specimens Examined	*Rock Type*	*Av. Value of Radioactivity*
3	Gabbro	0
1	Diorite dike	11
10	"Gabbro" inclusions	20
2	"Schist" inclusions	64
35	Tonalite	42
6	Schist	46
1	Aplite	52
8	Granodiorite	95
3	Granite	112
1	Pegmatite	220

various types of lavas from Lassen Volcanic National Park that the amount of radium increases with the alkalies, especially potassium.

Source of radioactivity.—More work is to be done by Wright on the radioactivity of the individual minerals of the Val Verde tonalite, but tentative conclusions may be reached concerning the constituent of the rock with which the radioactive material is associated. An examination of Figures 3 and 4 is sufficient to show that the data so far obtained indicate no correlation between the radium-thorium-actinium content of the tonalite and the percentage of biotite, quartz, or hornblende. A slight correlation exists between the low points in the plagioclase (Fig. 3) and radioactivity (Fig. 4) curves, but this may be fortuitous. Also, orthoclase is slightly more abundant in the rock near the contact where the radioactivity is higher. It is possible that the radioactive substances are contained largely in zircon,[7] which Wilson (1937) found to increase in amount toward the contact; in this case the amount of zircon in the tonalite may vary directly with the albite content of the plagioclase.

ORIGIN OF DARK INCLUSIONS IN TONALITE

The origin of the dark inclusions in the tonalite is an enigma not readily solved. If they are xenoliths, textural characteristics of the rocks from which they have been derived are lacking. The minerals, with exception of feldspar in some specimens, are the same as in the enclosing rock, but are in smaller grains, and the texture is granoblastic.

Twelve inclusions were examined for their mineralogic composition and radioactivity. The data obtained indicate that these are xenoliths of gabbro and schist. One specimen believed to have been originally

[7] Monazite is not included in the discussion because of its extreme rarity (Wilson, 1937, p. 126).

schist was taken from the center of a discoidal inclusion approximately 5 feet in diameter and one foot thick, located 400 yards from a schist-tonalite contact. This specimen contains 6 per cent orthoclase, which is unusual for any dark-colored rock in the district except schist. The plagioclase is oligoclase, which is the variety most common in the schist, but more sodic than that in the enclosing rock. The value for the radioactivity is 81, comparable to some of the higher values obtained from the schist. A second specimen believed to be a xenolith of schist was taken from an inclusion in the tonalite located near the western boundary of the district 20 feet from a schist-tonalite contact. A specimen of schist from the contact was taken for comparison. The inclusion is composed of 37 per cent oligoclase, 41 per cent quartz, 15 per cent biotite, and 7 per cent hornblende; the schist is composed of 38 per cent oligoclase, 25 per cent quartz, 17 per cent biotite, and 20 per cent hornblende. The value for the radioactivity of the inclusion is 46 and for the schist is 55. The grain size is the same in both. In view of the variability of the schist, this is a very close correlation. The composition of the plagioclase crystals in both inclusions is sufficient to exclude the possibility that these inclusions are segregations from the tonalite or xenoliths of gabbro.

The other ten inclusions studied are thought to have been derived from the gabbro. In the first place, the composition of the plagioclase is suggestive. Two specimens were taken from near the center of inclusions having a maximum diameter greater than 6 feet and a thickness of one foot or more. They were located near station 11915 (Fig. 2) on the line of the tunnel in an area in which inclusions compose 50 to 60 per cent of the rock. In both specimens some of the plagioclase grains have cores of labradorite or bytownite (An_{60-80}). The only rock in the district containing plagioclase as calcic as this is gabbro. From similar evidence Hurlbut (1935) concluded that the inclusions in the Bonsall tonalite were xenoliths of gabbro. If the other eight inclusions are gabbro xenoliths, as is thought, then the plagioclase in them, which is the same as that in the tonalite, has been changed by reaction with the tonalite magma. In the second place, the average value for the radioactivity of the ten inclusions is 20. This value is much lower than the average for the schist and it is not probable that the radioactivity of the schist would be lowered that much by reaction with the tonalite because the tonalite is 42. On the other hand, since the radioactivity of the gabbro is zero, reaction with the tonalite could change this value only by raising it. The ten inclusions are thus probably not xenoliths of schist. Some of them may have originated from other dark rocks not found on the surface, but inasmuch as gabbro is abundant in the district this is the most probable source.

The fine-grained texture of the inclusions may be related to the breakdown of the gabbroic texture when reacted upon by acidic magma. Joplin (1935) studied the problem of hybridization of basic rocks by acidic magma and found that in the first stage a granoblastic texture is produced in the basic rock and granular masses of pyroxene and crisscross flakes of biotite may form. During and following this stage large, highly poikilitic crystals of hornblende and biotite are developed. As Joplin points out, movement in the magma would disrupt the poikilitic crystals, producing a fine-grained aggregate. This same general process was observed in the Gavilan Peak gabbro. The gabbro at Gavilan Peak is uncontaminated by the tonalite and it contains augite and olivine, the former partly altered to uralite. The gabbro in the western and northern parts of the district is closely associated with the tonalite, and thin sections of this gabbro contain only hornblende and biotite as the mafic constituents. Moreover, a thin section of a specimen of gabbro taken from the contact with tonalite contains unusually large, highly poikilitic crystals of hornblende enclosing fine-grained andesine (Pl. 1, fig. 1). If a block of this gabbro were surrounded by tonalite magma and carried along with it, the movement probably would be sufficient to disrupt the poikilitic crystals and produce a fine-grained rock.

PETROFABRIC ANALYSES

GENERAL STATEMENT

The schist and tonalite in the Val Verde district were subjected to petrofabric analyses as an additional means of obtaining information on the geologic history. Of particular interest was the problem of the origin of flow structure or foliation in the tonalite. At the contact with the schist, the flow structure parallels the schistosity, striking in general northwest and dipping steeply northeast. Eastward from the contact with the schist, however, the dip of the foliation lessens, becoming approximately horizontal at the eastern end of the tunnel. The lineation in the tonalite parallels the dip of the gneiss planes. Is this orientation of the minerals in the tonalite entirely a product of magmatic flow, or may it be partly or wholly a result of post-magmatic deformation?

The method used was that developed by Sander (1930) and Schmidt (1932) and followed by other workers in the field. All megastructures such as contacts, folds, faults, joints, foliation, orientation of inclusions, and lineation in the tonalite were mapped. From the oriented hand specimens oriented thin sections were made and were examined by means of a Leitz universal stage. The microscopic data were recorded on an equal-area projection. The projection of the points was made from the lower hemisphere on the equatorial plane of the sphere.

PETROFABRIC ANALYSIS OF SCHIST

Axes of reference.—In accordance with the method of Sander, three axes of reference were set up for the oriented hand specimens of the schist. The *b*-axis is the tectonic axis, or that direction paralleling the axes of drag folds and small crinkles in the rock. The *a*-axis is that direction lying in the plane of schistosity or *ab* plane, 90° from *b*. The *c*-axis is normal to the *ab* plane. In the field the *b*-axis is horizontal or dips at an angle less than 10° to the northwest and is directed in a general north to northwest direction. The *a*-axis, therefore, approximately parallels the dip of the beds, which is in most places steep and to the east.

Fabric diagrams.—Oriented thin sections were made of four specimens of the schist, and optic axes of quartz grains and poles of basal planes of biotite were plotted. Figure 6 shows the results obtained on two of the specimens in thin sections cut normal to the *b*-axis. In the diagrams, *c* and *b* are horizontal and *a* vertical, approximately as in the field. As shown in diagrams 1 and 3, the poles of the basal planes of biotite are concentrated about *c*. Inasmuch as the foliation, or *ab* plane in the rock, is outlined in hand specimen by basal planes of biotite flakes, this orientation is exactly as would be expected. Biotite is not only concentrated at *c*, but also forms a nearly complete girdle about *b*. The optic axes of quartz grains, as shown in diagrams 2 and 4 also describe girdles about *b*. Several maxima are present in each diagram, with none appearing at *a*.

Joints and faults.—In general joints are not conspicuous in the schist, but in the body of schist half a mile south of the western end of the tunnel, well developed joints are common striking approximately at right angles to the strike of the schist and dipping vertically or steeply to the south. These are thus approximately parallel to the *ac* plane of the schist fabric and undoubtedly represent the *ac* joints of Sander.

The main fault in the district lies along the western border of the schist and parallels the schistosity.

Interpretation.—Inasmuch as the schistosity parallels the bedding, the orientation of the biotite flakes may be largely a residual structure. There may, therefore, have been no rotation of biotite grains into the *ab* plane during deformation. Gliding within the biotite parallel to the cleavage may have occurred during movement of the rock, but this is not proven. The preferred orientation of quartz resulted from movement in the rock and might be explained by rotation of the grains in a plane normal to *b*, and by gliding along crystallographic planes. Inasmuch as only one *s*-surface is visible in hand specimens and only one is appar-

ent in the biotite diagrams, the various quartz maxima are not readily explained as being due to the presence of two or more *s*-surfaces. By assuming gliding in quartz parallel to rhombohedrons during shearing parallel to the *ab* plane, the occurrence of the quartz maxima may be

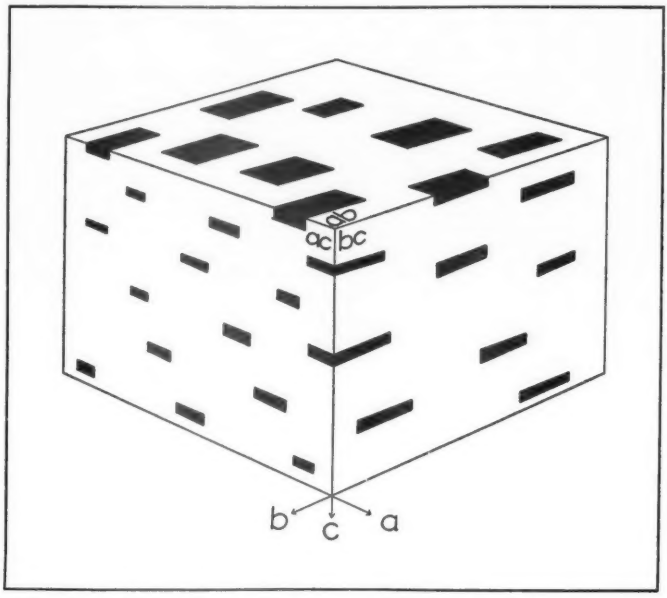

FIGURE 5.—*Diagrams illustrating relation of axes of reference to dimensional orientation of grains in tonalite*

explained reasonably well. But this assumption may not be valid. Recent experiments by Griggs and Bell (1938) indicate that a translation gliding mechanism is not present in quartz, and that during rock deformation quartz is fractured into needles whose elongation direction and bounding faces bear a relation to the crystal lattice. The needles are oriented in the *a* fabric direction, with one of their bounding planes in the *ab* plane. Later crystallization is assumed to obliterate traces of the needles. On the basis of this theory, only one shear plane in the rock is necessary.

PETROFABRIC ANALYSIS OF TONALITE

Axes of reference.—Axes of reference used for the tonalite have in general the same significance as for the schist despite the fact that one is an igneous and the other a metamorphic rock. As discussed by

Sander and Felkel (1929), the same types of motion may occur in solidifying magmas as in tectonically deformed rocks, and the Val Verde tonalite is an igneous body for which this apparently holds true. The

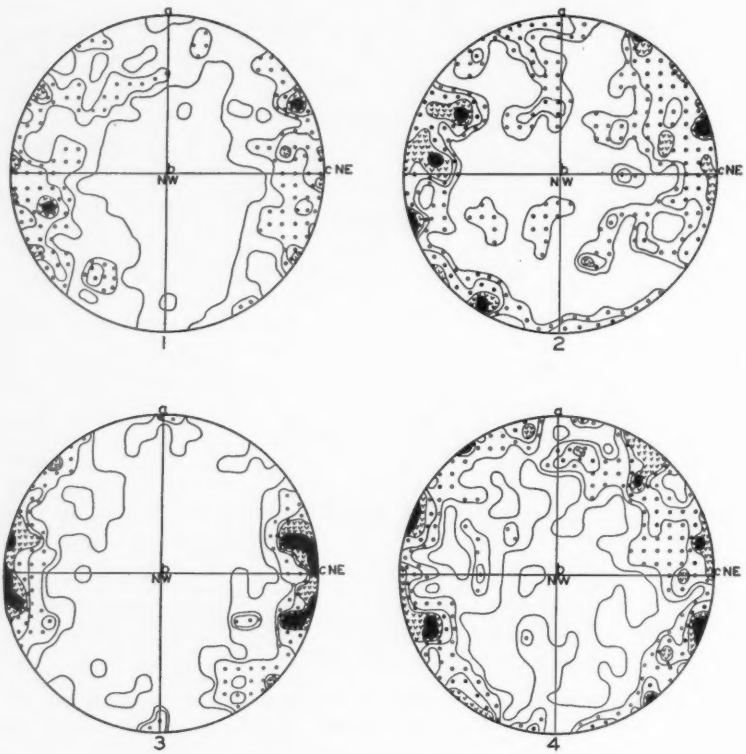

FIGURE 6.—*Fabric diagrams of schist*

(1) Poles of basal planes of 308 biolite grains in sp. 12012 of quartz-biotite schist. Contours 5, 4, 3, 2, 1%.
(2) Optic axes of 183 quartz grains in sp. 12012 of quartz-biotite schist. Contours 5, 4, 3, 2, 1%.
(3) Poles of basal planes of 343 biotite grains in sp. 12011 of quartz-biotite schist. Contours 5, 4, 3, 2, 1%.
(4) Optic axes of 319 quartz grains in sp. 12011 of quartz-biotite schist. Contours 5, 4, 3, 2, 1%.

direction of movement in the tonalite, as shown by fabric analyses, lies in a plane normal to the lineation or direction of elongation in the rock. The lineation in the tonalite, therefore, is labeled *b*, signifying the axis of intersecting surfaces of movement or the axis about which grains

have been rotated. In the field, *b* parallels the dip of the foliation, which is steep at the schist contact, but gradually approaches the horizontal eastward. The *a*-axis lies in the plane of foliation and is normal to *b;*

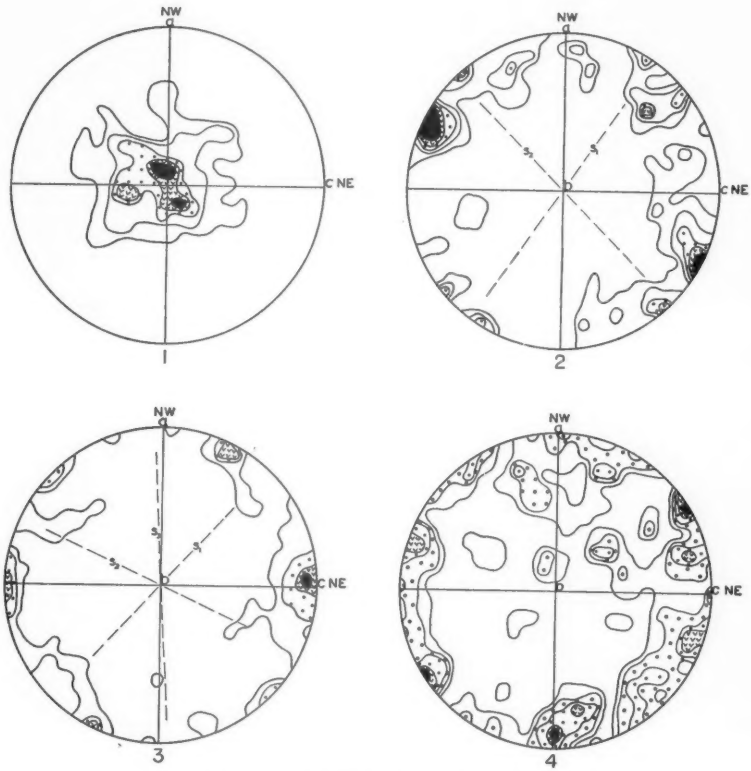

FIGURE 7.—*Fabric diagrams of tonalite*

(1) *a*-axes of 106 andesine grains in sp. 11868 of tonalite. Contours 10, 8, 6, 4, 2%.
(2) Poles of 010 faces of 106 andesine grains in sp. 11868 of tonalite. Contours 5, 4, 3, 2, 1%.
(3) Poles of basal planes of 88 biotite grains in sp. 11868 of tonalite. Contours 4, 3, 2, 1%.
(4) Optic axes of 185 quartz grains in sp. 11868 of tonalite. Contours 5, 4, 3, 2, 1%.

a is thus approximately horizontal and trends northwest. The *c*-axis is normal to the *ab* plane, or plane of foliation. Figure 5 and Plate 3 illustrate the scheme of orientation and the latter gives also an idea as to the degree to which the minerals are oriented in some of the tonalite specimens.

Fabric diagrams.—Nine specimens of tonalite from the tunnel, distributed between stations 11734 and 12022, were examined. Typical diagrams obtained from three specimens of tonalite and one of an inclu-

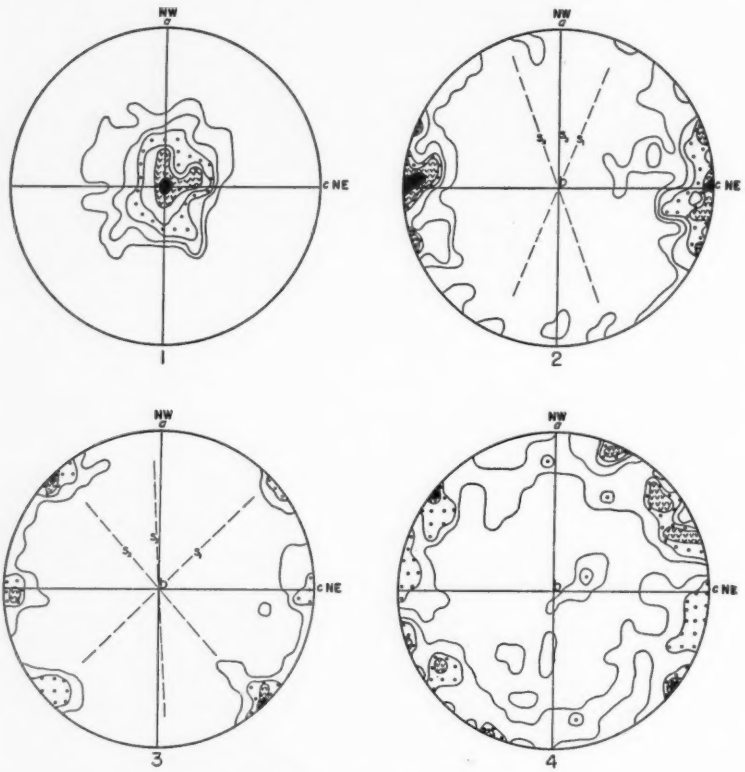

FIGURE 8.—*Fabric diagrams of tonalite*

(1) *a*-axes of 130 andesine grains in sp. 11870 of tonalite. Contours 12, 10, 8, 6, 4, 2%.
(2) Poles of 010 faces of 130 andesine grains in sp. 11870 of tonalite. Contours 10, 8, 6, 4, 2%.
(3) Poles of basal planes of 69 biotite grains in sp. 11870 of tonalite. Contours 4, 3, 2, 1%.
(4) Optic axes of 163 quartz grains in sp. 11870 of tonalite. Contours 4, 3, 2, 1%.

sion are shown in Figures 7, 8, and 9. Since lineation is not evident in the inclusions, the oriented section from the inclusion was cut normal to the linear direction in the adjoining tonalite. The positions of *s*-surfaces are labeled S_1, S_2, and S_3. These lines have been drawn parallel to concentrations of the (010) face of plagioclase and the basal plane of biotite.

The lineation in the tonalite is determined by subparallel arrangement of prisms of hornblende and grains of plagioclase elongated parallel to the *a* crystallographic axis. The degree to which lineation in the rock

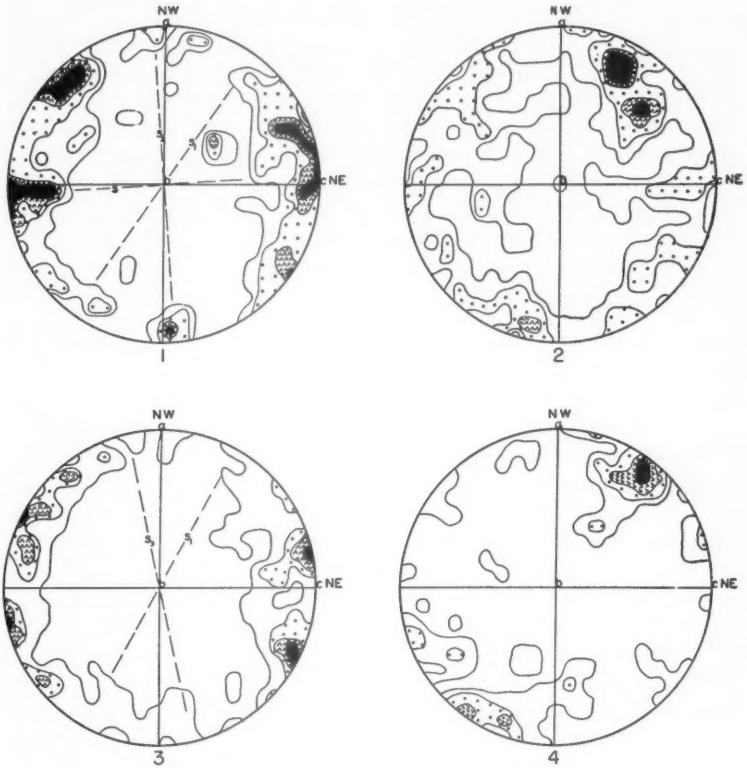

FIGURE 9.—*Fabric diagrams of tonalite and an inclusion*

(1) Poles of basal planes of 165 biotite grains in sp. 11982 of tonalite. Contours 4, 3, 2, 1%.
(2) Optic axes of 260 quartz grains in sp. 11982 of tonalite. Contours 4, 3, 2, 1%.
(3) Poles of basal planes of 143 biotite grains in sp. 11930 of an inclusion in tonalite. Contours 6, 4½, 3, 1½%.
(4) Optic axes of 108 quartz grains in sp. 11930 of an inclusion in tonalite. Contours 8, 6, 4, 2%.

is developed was determined by plotting the *a*-axis of the plagioclase in sections cut normal to the linear direction in the rock. Figure 7, (1); and Figure 8, (1) illustrate this.

Figure 7, (2) and Figure 8, (2) are plots of the poles of the (010) faces of plagioclase grains in the same two specimens. A discontinuous

girdle is present in the plane normal to *b*, and in this girdle conspicuous concentrations of (010) faces occur in two intersecting planes labeled S_1 and S_2, almost symmetrically arranged with respect to the *ab* plane.

The quartz diagrams (Fig. 7, (4), Fig. 8, (4), and Fig. 9, (2)) possess a girdle of quartz optic axes in the plane normal to the *b* tectonic axis, and within this girdle rather strong maxima occur, somewhat as in the schist diagrams.[8] Figure 9, (4), a plot of the quartz grains in an inclusion in the tonalite, resembles Fig. 9, (2), but differs from it in having a less distinct and continuous girdle.

The biotite diagrams (Fig. 7, (3), Fig. 8, (3), and Fig. 9, (1) and (3)) of tonalite and an inclusion are from sections cut normal to *b*. Incomplete girdles occur in all diagrams, and besides a common concentration of poles of the basal planes at *c*, concentrations occur at other points in the *ab* plane showing the presence of *s*-surfaces inclined to *ab*, similar to those appearing in the plagioclase diagrams.

Interpretation.—Fabric diagrams of quartz, biotite, and plagioclase in the tonalite are interpreted to mean that some movement in the tonalite was produced by rotation of mineral grains in a plane normal to the *b*-axis. The positions of the maxima of poles of the basal plane of biotite and (010) face of plagioclase indicate that three surfaces of movement, labeled S_1, S_2, and S_3 on the diagrams, were important during deformation. If gliding in quartz occurred, and if the gliding was parallel to prism faces, similar shearing directions are indicated by the quartz maxima.

The main object of this study has been to uncover evidence on the origin of the gneissoid structure in the tonalite and thus to obtain a more complete picture of the history of the intrusive. Three possibilities exist: (1) the structure was developed during the magmatic stage of the tonalite; (2) the structure was developed after complete solidification of the rock; and (3) the structure is a result of stresses operative on the rock both during and after solidification.

(1) In support of the hypothesis that the foliation is entirely a result of magmatic flow the following points may be listed[9]: (a) The texture of the tonalite is essentially that of an igneous rock. Evidence of metamorphism is slight. (b) The attitude of the foliation and the direction of lineation are those normally to be expected were the foliation the result of primary flow (Balk, 1937). That is, at contacts with country rock the foliation is parallel to the contact and the linear direction is

[8] Hurlbut (1935) and Pabst (1936) found no orientation of quartz in either the Bonsall tonalite, southern California, or in the intrusive rocks of the Sierra Nevada.

[9] Joint and fault patterns might be of use in distinguishing primary and secondary foliation, but the writer has not been able to use the fault and joint patterns mapped in the tunnel as evidence for any theory on the origin of the flow structure.

parallel to the dip of the contact. (c) Inclusions are oriented parallel to the foliation, and it is most reasonable, especially in view of their large size and gradational borders in some instances, to assume that the inclusions were oriented thus before complete solidification of the tonalite. The foliation, therefore, must have been developed, at least partly, before complete solidification.

The only evidence opposing this first hypothesis is that furnished by the petrofabric diagrams. The intersecting surfaces in the rock containing basal planes of biotite and (010) faces of plagioclase are not readily explained, the writer believes, as being the result of primary flow. A wavering of the flow lines is to be expected, but the development of one or more definite surfaces inclined to the foliation is a different matter. Moreover, the optic axes of quartz are oriented in a similar manner. The quartz grains are irregular and are not flattened or elongated in any direction bearing a constant relation to the optic axis. The arrangement of the quartz, therefore, is undoubtedly not the result of fluidal movement in the magma. Furthermore, the optic axes of the quartz grains form a girdle normal to the linear direction in the rock, and, from studies of metamorphic rocks (Sander, 1930), this is believed to mean rotation of the grains in a plane normal to the linear direction. If the orientation of the quartz is a result of post-magmatic deformation, then, since the orientation of the biotite and plagioclase can be correlated with the quartz orientation pattern, may not the arrangement of all of these minerals be the result of deformation following solidification of the tonalite?

(2) Evidence favoring the hypothesis that foliation was developed following solidification of the tonalite is as follows: (a) Orientation patterns of the minerals can be better explained by deformation following solidification of the rock. Much is yet unknown about the process of orientation of minerals during metamorphism, but patterns similar to those presented herein have been found commonly in metamorphic rocks (Sander, 1930). In the diagrams of quartz, a girdle has developed normal to *b*, and distinct maxima occur within this girdle. The actual process by which quartz was oriented is not certain, but the grains undoubtedly rotated about an axis and may have been fractured into needles which were mostly aligned in the shear planes (Griggs and Bell, 1938, p. 1737-1744). The "segmented" quartz grains described on an earlier page may represent incompletely recrystallized needles. A similar movement on the part of the plagioclase and biotite is shown by presence in the diagrams of *s*-surfaces inclined to the *ab* plane, as well as girdles. This orientation pattern is believed to be an example of flattening structure (German: *Plättung*) (Sander, 1930, p. 220), which is a dimensional

orientation of the grain fabric. In most of the biotite and plagioclase diagrams a strong concentration of the poles of basal planes and (010) faces, respectively, occurs at *c;* or in other words, a large proportion of the grains lies in the flattening surface. This surface represents an *s*-surface developed at a different time, and may be the original *s*-surface in the rock. The elongation of the plagioclase and hornblende parallel to *b* is also compatible with this hypothesis, for in tectonites, the direction of elongation is most commonly normal to the girdle of quartz and biotite. (b) The inclusions possess the same type of mineral orientation as the tonalite, indicating that both inclusions and the host rock were subjected to the same type of movement and that orientation in each was developed at the same time. The fact that the inclusions do not show a preferred orientation within the plane of foliation but do have a *b*-axis that is parallel to the *b*-axis of the tonalite is additional evidence in favor of the post-consolidation origin not only of the quartz fabric but also of the biotite fabric. (c) In support of the theory that minerals in some igneous rocks have been oriented by post-magmatic deformation, we have results of petrofabric studies made on other intrusive bodies. Johs (1933) examined a quartz porphyry possessing a linear structure parallel to that in a mica schist intruded by the porphyry. A quartz girdle normal to the linear direction is present in both rocks. The direction of lineation in the porphyry is not that to be expected were the mineral orientation produced by magmatic flow. He concluded that deformation, probably during the final stages of solidification, had played some part in the orientation of the minerals of the quartz porphyry. A granite, which is largely massive, was studied by Maroschek (1935). Even though no linear direction in the rock was megascopically observable, girdles of quartz were found to occur, and these correlated with the joint pattern in the rock as if the granite were a true tectonite. In some of the gneisses studied by Sahlstein (1935), girdles of quartz were found oriented at right angles to the foliation and to the linear direction in the rock, a case similar to that of the Val Verde tonalite.

Opposed to the hypothesis that the arrrangement of the minerals in the tonalite was produced entirely following solidification is evidence furnished by the parallel orientation of flattened inclusions, as discussed under (1.,c).

(3) A third hypothesis explains the structure as having been produced by a combination of flow before, and deformation following, solidification. A time interval between the period of flow of the magma and the period of deformation of the crystallized tonalite is not required. The earlier stage may have graded into the later as the rock increased

in crystallinity. This combination of the first and second hypotheses is necessary to explain all the phenomena. On the basis of the foregoing discussion it is believed that the foliation and inclusions received their orientation during flow of the partly crystallized magma. Some reorientation of mineral grains occurred, however, as a result of shearing stresses acting on the solid rock. How significant post-magmatic rearrangement of grains may have been is not clear. Rotation of quartz, biotite, and plagioclase crystals in a plane normal to the linear direction in the rock seems to have occurred, but was the *a*-axis of plagioclase (the linear direction in the rock) appreciably shifted in direction, or is the present linear direction the original direction of elongation of the magma? The writer believes there is a possibility that the linear direction is a structure impressed upon the rock during the final stages of, or following, solidification and is possibly independent of the original direction of elongation of the magma. But the fact that the texture of the tonalite is dominantly igneous rather than crystalloblastic is evidence for believing that post-magmatic deformation was not sufficiently intense to disturb primary lineation in the rock.

If we assume that lineation in the tonalite is a primary feature, then development of orientation of the minerals may be viewed from two standpoints depending on direction of intrusion of the magma: (a) The magma was intruded from the northeast up the dip of the schist. Long directions of crystals were arranged parallel to the direction of motion of the magma and flat surfaces were arranged parallel to the contacts with country rock. Deformative stresses, which may have been active in the region during intrusion, took effect on the tonalite after it possessed sufficient rigidity to transmit differential stress. The stresses were of such a nature that the rock tended to be sheared parallel to the contact and in the direction of the strike, with the resulting rotation of the minerals. The existing foliation in the rock influenced the attitude of the plane of rotation of the minerals to the extent that this plane of rotation was maintained normal to the foliation. The foliation was transformed into a flattening structure by rearrangement of the mineral grains. (b) The magma was pushed in either from the northwest or southeast in a direction parallel to the strike of the country rock. The actual elongation of the magma, however, was up dip, so that a linear direction parallel to the dip of the country rock was developed. That is, a linear direction was developed at right angles to the direction of intrusion of the magma. The same forces that were instrumental in emplacing the magma produced the later rotation of mineral grains about the linear direction in the rock. By thus assuming a horizontal direction of intrusion parallel to the later translation direction in the solid state, the

coincidence of a later stress being applied normal to an earlier force does not have to be invoked. Furthermore, a horizontal direction of intrusion might explain the small folds and crinkles with axes parallel to the dip of the schist which commonly occur in the schist near the contact and seem to be related to intrusion of the tonalite.

The writer concludes that the foliation developed in the tonalite as a combined result of flow before complete solidification, and of post-magmatic deformation. The *s*-surfaces and girdles may have been produced during the late stages in the crystallization of the tonalite, but it is difficult to understand how *s*-surfaces could have developed if the rock were not essentially solid. The actual amount of post-magmatic movement may have been small. The linear direction now visible within the tonalite probably represents the original direction of elongation of the magma and may be parallel or normal to the direction of intrusion.

The fabric pattern of the tonalite is not present in the schist specimens which were examined. The *ab* plane in the schist is parallel to that in the tonalite, but the *a*-axis in the schist is parallel to the *b*-axis in the tonalite. If the movement which produced the mineral orientation in the tonalite had similarly occurred in the schist, at least a semblance of a girdle normal to *a* should appear in the diagrams of the schist. Since this is not the case, it is concluded that movement in the tonalite was localized, that is, confined to the tonalite. The movement probably occurred while the tonalite was more plastic than the schist.

CONCLUSIONS

The results of micrometric and radioactivity measurements of the Val Verde tonalite may be summarized as follows. (1) The linear method of Rosiwal is accurate to within about one per cent for an individual constituent, in the sense in which per cent is used here. (2) A thin section greater than 300 sq. mm. samples a hand specimen of the tonalite within the limits of error of the measuring method. (3) A hand specimen samples a body of tonalite six feet in diameter probably with an error of less than 3 per cent for one mineral. (4) Micrometric and radioactivity measurements made on samples of tonalite taken along a line extending from the border of the tonalite intrusive to a point 5 miles out in the intrusive have shown that: (a) none of the minerals studied increases or decreases regularly in amount along the line, although orthoclase is slightly more abundant near the border: (b) the albite content of the plagioclase and radioactivity of the tonalite both increase toward the border. Heavy mineral studies have shown that zircon also increases in abundance toward the border. If radioactive substances are contained in zircon, then this mineral may vary in amount directly as the albite

content of the plagioclase. (5) The more acidic border of the tonalite intrusive is believed to be due to assimilation of quartz-biotite schist. (6) On the basis of mineralogic and radioactivity data it is believed that the dark, fine-grained inclusions in the tonalite are xenoliths of gabbro and quartz-biotite schist.

The geologic history of the district may be divided into five stages. (1) During the Triassic period arenaceous sediments were deposited. (2) Later in the Mesozoic, after burial and consolidation of the sedimentary series, the minerals were recrystallized and the beds folded along axes striking northwest. Shearing parallel to the beds accompanied the folding, rotating the grains about the tectonic axis. (3) Following the folding of the schist and before the Lower Cretaceous, partly crystalline magma entered the schist and assumed a flow structure parallel to its contacts. Assimilation of the schist seems to have been important during emplacement of the intrusive. (4) Either before the rock had become entirely crystalline, or following this stage, horizontal movement parallel to the foliation occurred in the rock, effecting a new orientation for some or all of the minerals in the tonalite. Movement along the strike slip faults that occur within and bordering the Perris Block may have been associated with the horizontal movement in the tonalite. (5) The history of the Perris Block during the Tertiary period is incompletely known and these studies give us no additional information. At present the block appears to be down-dropped with respect to the districts to the north, west, and east, and it is also apparently tilted downward toward the east.

WORKS TO WHICH REFERENCE IS MADE

Balk, Robert (1937) *Structural behavior of igneous rocks,* Geol. Soc. Am., Mem. 5. 177 pages.

Dudley, P. H. (1935) *Geology of a portion of the Perris Block, southern California,* Calif. Jour. Mines and Geol., vol. 31, p. 487-505.

——— (1936) *Physiographic history of a portion of the Perris Block, southern California,* Jour. Geol., vol. 44, p. 358-379.

English, W. A. (1926) *Geology and oil resources of the Puente Hills region of California,* U. S. Geol. Survey, Bull. 768, 110 pages.

Evans, R. D., and Williams, Howell (1935) *The radium content of lavas from Lassen Volcanic National Park, California,* Am. Jour. Sci., 5th ser., vol. 29, p. 441-452.

Fairbanks, H. W. (1892) Calif. Min. Bur., 11th Ann. Report, p. 106-107 and 111-113.

Fletcher, A. L. (1911) *The radioactivity of the Leinster granite,* Philos. Mag., vol. 21, p. 102-111.

Goodyear, W. A. (1888) Calif. Min. Bur., 8th Ann. Rept., p. 526-527.

Griggs, David, and Bell, J. F. (1938) *Experiments bearing on the orientation of quartz in deformed rocks,* Geol. Soc. Am., Bull., vol. 49, p. 1723-1746.

Hirschi, H. (1927) *Radioaktivität der Intrusivegesteine des Aarmassivs,* Schweiz. Min. u. pet. Mitt., vol. 7, p. 98-114.

Hurlbut, C. S., Jr. (1935) *Dark inclusions in a tonalite of southern California,* Am. Mineral., vol. 20, p. 609-631.

Johs, Max (1933) *Der Granitporphyr von Thal-Heiligenstein im Thuringer Wald,* Tscherm. Min. u. pet. Mitt., vol. 43, p. 283-318.

Joplin, G. A. (1935) *Origin of basic xenoliths in plutonic rocks,* Geol. Mag., vol. 72, p. 227-235.

Larsen, E. S., and Miller, F. S. (1935) *The Rosiwal method and the modal determination of rocks,* Am. Mineral., vol. 20, p. 260-273.

Maroschek, E. F. (1935) *Beiträge zur Kenntnis des Granites von Mauthausen in Oberösterreich,* Tscher. Min. Pet. Mitt., vol. 43, p. 375-405.

Pabst, Adolf (1936) *Orientation of minerals in "autoliths",* Am. Mineral., vol. 21, p. 68.

Piggot, C. S., and Merwin, H. E. (1932) *Radium in rocks: IV. Location and association of radium in igneous rocks,* Am. Jour. Sci., 5th ser., vol. 23, p. 49-56.

Ransome, F. L. (1932) *Final geological report on the Bernasconi and Val Verde tunnel lines, Colorado River Aqueduct,* unpublished.

Sahlstein, Th. G. (1935) *Zur Regelung der Gesteine im Kristallin von Liverpool-Land in Ostgrönland,* Meddelelser om Grönland, vol. 95, p. 1-23.

Sander, Bruno (1930) *Gefuegekunde der Gesteine,* Julius Springer, Vienna. 352 pages.

——— ———, and Felkel, E. (1929) *Zur tektonischen Analyse von Schmelztektoniten,* Sitzungsber. der Heidelberger Akademie der Wissensch., Math.-Nat. Kl., vol. 13, p. 1-40.

Schmidt, Walter (1932) *Tektonik und Verformungslehre,* Borntraeger, Berlin. 208 pages.

Strutt, R. J. (Lord Rayleigh) (1905) *On the distribution of radium in the earth's crust, and on the earth's internal heat,* Royal Soc. of London, Pr., vol. 77, p. 472-485.

Waring, G. A. (1919) *Ground water in the San Jacinto and Temecula basins, California,* U. S. Geol. Survey, Water Supply Paper 429, p. 15.

Waters, J. W. (1909) *Radioactive minerals in common rocks,* Philos. Mag., vol. 18, p. 677-679.

——— (1910) *Radioactive minerals in common rocks,* Philos. Mag., vol. 19, p. 903-904.

Wilson, R. W. (1937) *Heavy accessory minerals of the Val Verde tonalite,* Am. Mineral., vol. 22, p. 122-132.

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CAMBRIAN FORMATIONS OF SOUTHWESTERN ALBERTA AND SOUTHEASTERN BRITISH COLUMBIA

BY CHARLES DEISS

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ABSTRACT

Field studies since 1931 have shown that: published type sections in Alberta and British Columbia were not usable for accurate correlation; genera restricted to widely separated stratigraphic horizons were lumped together in published faunal lists; formations, as defined, were not mappable units; and the Middle-Lower Cambrian boundary was incorrectly located.

Three sections were remeasured in 1938: Castle Mountain and Ptarmigan Peak in Alberta, and Ross Lake in British Columbia. The results are:

Emended definitions are given of the Lower Cambrian Fort Mountain, St. Piran, and Mount Whyte formations, and of the Middle Cambrian Ptarmigan, Cathedral, and Eldon formations. Fossils are zoned in each section, and revised faunal lists are given for each formation in which fossils were found. The Lake Louise shale is placed as a member of the Fort Mountain sandstone.

The new basal Upper Cambrian Pika formation is defined. Discovery of a pre-*Cedaria* fauna in these rocks lowers the Upper-Middle Cambrian boundary and furnished more precise evidence for its stratigraphic position.

The Mount Whyte formation of earlier writers is both Middle and Lower Cambrian in age. The upper 58 feet contains *Kochaspis* cf. *cecinna* and *K.* cf. *gogensis* and is lithologically identical with Ptarmigan limestone. Consequently, the Middle-Lower Cambrian boundary must be drawn just below the *Kochaspis cecinna* zone and between the emended Mount Whyte and Ptarmigan formations.

The Lower and Middle Cambrian formations are correlated tentatively with those in northwestern Montana, the House Range in Utah, and the Highland Range in Nevada.

INTRODUCTION

THE PROBLEM

When the Cambrian sections in northwestern Montana were measured in 1932 (Deiss, 1933, p. 4-33) the fact was not realized that the older type sections of Peale (1893, p. 20-25) and Weed (1900) in central Montana needed revision. Field work in 1933 and especially the study of new collections of trilobites during 1932 to 1934 in comparison with published types in the United States National Museum necessitated remeasuring the type Cambrian sections in central Montana, redefining the older standard and often quoted formations, revising and clarifying the stratigraphic nomenclature, and accurately locating newly collected faunas in their proper stratigraphic positions. Only after the completion of that revision (Deiss, 1936, p. 1268-1341) could the new Cambrian formations in northwestern Montana be accurately defined and the faunas correlated with those in central Montana and Yellowstone National Park (Deiss, 1939, p. 34-58).

In attempting to correlate the Middle Cambrian formations and sections of Montana with the type sections to the north and south in the Cordilleran area, numerous incongruities and contradictions were encountered in publications. The errors in the literature necessitated remeasuring several of the more strategic older type sections in Nevada, Utah, and Wyoming to the south, and in Alberta and British Columbia to the north.

Four southern sections were remeasured in 1937: the Highland Range in Nevada, the House Range and Blacksmith Fork in Utah, and the Wind River Canyon in Wyoming (Deiss, 1938, p. 1067-1168). The field and laboratory studies in 1937 added a new problem in connection with the Cambrian rocks of Alberta and British Columbia—that of the exact position of the Middle-Lower Cambrian boundary.

Thus, by 1938, the field studies in the Cordilleran area of the United States had shown that: (1) The published type sections in Alberta and British Columbia were not usable for accurate correlation, (2) the formation boundaries were drawn at different stratigraphic horizons in the

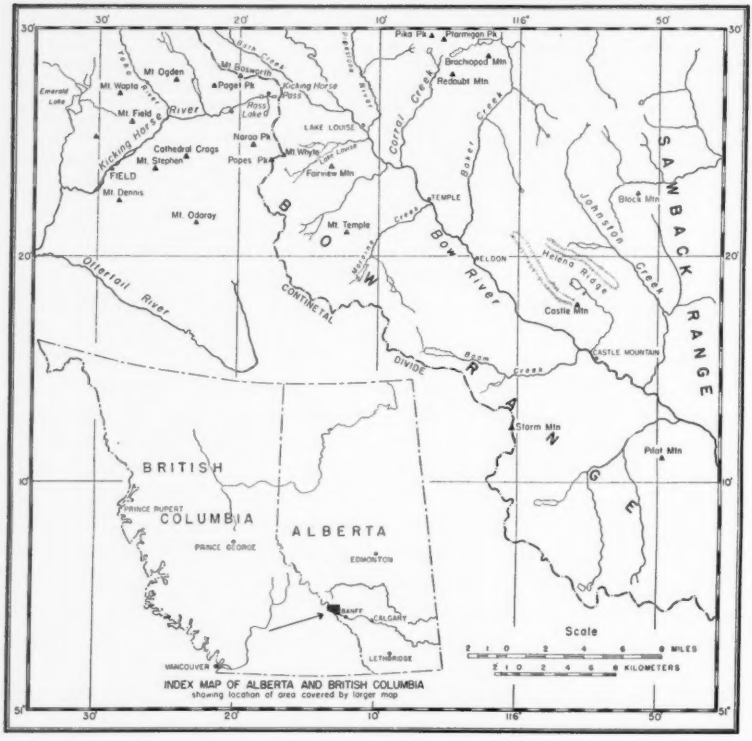

FIGURE 1.—*Map of area described*

different sections, (3) the exact stratigraphic horizons, and therefore the relative ages of many genera and species were unknown, (4) genera now known to be restricted to one of several widely separated stratigraphic horizons were lumped together in the published faunal lists, (5) some of the formations as defined were not mappable units, and (6) the Middle-Lower Cambrian boundary was incorrectly located.

As the first step in the attack on these six problems the writer remeasured three sections in the field season of 1938: Castle Mountain and Ptarmigan Peak in Banff National Park, southwestern Alberta, and

Ross Lake Cirque in Yoho National Park, southeastern British Columbia. The results of that field work comprise this report.

LOCATION OF AREA

The area investigated is in the southern part of the Canadian provinces, Alberta and British Columbia, between the parallels 51° 15′ and 51° 30′ north latitude and the meridians 115° 54′ and 116° 20′ west longitude. The three sections form an obtuse triangle whose base, approximately 20 miles in length, trends N. 60° W., whose short side, 10 miles in length, trends N. 70° E., and whose intermediate side, 15 miles in length, trends N. 30° W. Castle Mountain lies at the southeast angle in an unnamed range of mountains; Ptarmigan Peak lies at the northeast angle in the Slate Mountains; and Ross Lake Cirque lies at the northwest angle on the north end of the Bow Range (Fig. 1).

ACKNOWLEDGMENTS

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SUMMARY OF PREVIOUS WORK

Cambrian fossils in this part of Alberta and British Columbia were first discovered in 1884 by L. M. Lambe of the Geological Survey of Canada (McConnell, 1889, p. 23) who found them in talus at the base of Mount Stephen.

Cambrian strata in Alberta and British Columbia probably were first described by McConnell (1887, p. 24D-30D), who measured a generalized section in the vicinity of Castle Mountain and subdivided the rocks into the Castle Mountain group (limestone and dolomites) and the underlying

Bow River group (argillites, sandstones, quartzites, and conglomerates). The Castle Mountain group included strata belonging to the Lower, Middle, and Upper Cambrian, and Ordovician. The Bow River group contained strata now assigned to the Paleozoic (Lower Cambrian) and Proterozoic (Beltian) eras.

Cambrian fossils from the area were first described and figured by Rominger (1887, p. 12-19; Pl. 1). Although Rominger did not know the stratigraphic position or age of his fossils, they have since been shown to belong to the *Ogygopsis* shale member of the Stephen formation.

The following year Walcott (1888, p. 162-166) emended Rominger's generic identifications, assigned the fauna to the Middle Cambrian, and correlated it with part of the Highland Range section in Nevada.

In 1889 McConnell (1889, p. 22-25) discussed the stratigraphy in the vicinity of Mount Stephen and assigned the beds containing the fauna described by Rominger to the Bow River group, thus raising the boundary between the Bow River and Castle Mountain groups and changing their meaning.

In 1890 Walcott (1890, p. 550, 571-572) discussed McConnell's section (McConnell, 1887, p. 28D-30D) and suggested that the argillites and sandstones of the Bow River group may be correlative with the Algonkian strata of the Wasatch Range in Utah.

In 1901 Dawson (1901, p. 64-68) reviewed the existing work on the Cambrian rocks in the vicinity of Castle Mountain and Mount Stephen, summarized the descriptions of the Bow River series [1] and Castle Mountain group, and compared the eastern and western Cambrian deposits in the Canadian Rockies.

Additional field work was not done on the Cambrian rocks in the area during the following 6 years. In the summer of 1907 Walcott measured the Castle Mountain, Mount Bosworth, and the lower part of the Mount Whyte and Mount Stephen sections. In 1908 Walcott (1908a, p. 1-5) subdivided the Cambrian rocks into 10 formations and defined them. He assigned the Fairview, Lake Louise, St. Piran, and Mount Whyte [2] formations to the Lower Cambrian, the Cathedral, Stephen, and Eldon to the Middle Cambrian, and the Bosworth, Paget, and Sherbrooke to the Upper Cambrian. Later in 1908 Walcott (1908b, p. 15-52) published descriptions of four new species of trilobites from these rocks, a detailed section which he measured on Mount Bosworth (1908c, p. 204-217), and another estimated section on Mount Stephen (1908d, p. 232-248). In the former section Walcott assigned the rocks which form the top of

[1] The terms "Bow River series" and "Bow River group" were used interchangeably by the early writers.

[2] In his comparative sections Walcott (1908a, p. 2) assigned the Mount Whyte formation to the base of the Middle Cambrian.

Paget Peak to the Ordovician. In the Stephen section he gave lists of fossils for the Stephen and Whyte [*sic*] formations and figured some species from the *Ogygopsis* shale.

In 1910 Walcott (1910a, p. 425-427) described the basal Cambrian conglomerate and the unconformable relationships between the Cambrian and pre-Cambrian rocks in the vicinity of Bow Valley, and particularly near Ptarmigan Peak and Redoubt Mountain.

The next year Allan and Burling found Upper Cambrian fossils in the rocks at the top of Paget Peak (Walcott, 1912a, p. 229) and thus established the Upper Cambrian (instead of Ordovician) age of the rocks which form the top of the Mount Bosworth section of Walcott. Later in 1911 Alan (1912, p. 178) published the first measured section of the Upper Cambrian rocks which lie to the west of the Bow Range and which he defined the next year as the Chancellor and Ottertail formations (Allan, 1912, p. 179-180).

In 1912 Walcott (1912a, p. 229-231) discussed the Cambro-Ordovician boundary in British Columbia and the relationships of Allan's Chancellor and Ottertail formations to the Sherbrooke, Paget, and Bosworth formations on Mount Bosworth. Later the same year Walcott (1912b, p. 131, footnote a) substituted the name Fort Mountain sandstone for the preempted name Fairview formation for the basal sandstones of the Lower Cambrian series in Alberta and southeastern British Columbia in the vicinity of Redoubt (old Fort Mountain) Mountain and Mount Bosworth.

In 1913 Allan (1913, Route map between Banff and Golden) published the first geologic map of the Cambrian rocks between Castle Mountain and Kicking Horse Pass and divided the rocks into three unnamed cartographic units corresponding to the Lower, Middle, and Upper Cambrian. Allan (1913, p. 174-179) also summarized the Cambrian stratigraphy of the area and assigned the strata to Walcott's formations. In the same year Walcott (1913, p. 342-343) published his section measured in the Robson Peak district and compared it with the Mount Bosworth section.

Several important contributions to the Cambrian geology in the area were made in 1914. Allan (1914a, map 142a) published the only comprehensive geologic map yet made of the Cambrian rocks in the vicinity of Field, British Columbia. In that paper Allan also quoted his descriptions of the Chancellor and Ottertail formations in the Field map area. These have not been found east of the Continental Divide. In the same year the summary report (Allan, 1914b, p. 165-176) was published of Allan's field work in 1912 along the Canadian Pacific Railway between Banff and Golden, upon which was based part of the Transcontinental Excursion Guidebook 28.

An important paper dealing with the Cambrian stratigraphy of the area was published by Burling (1914, p. 93-129). He correctly assigned the *Albertella* fauna to the Middle instead of Lower Cambrian and drew the Middle-Lower Cambrian boundary at the base of the *Albertella* zone. Burling also correlated the *Albertella* beds in Canada with the Pioche shale of Nevada.

The other paper published in 1914 was a guide to the geology of the Canadian National Parks. Camsell (1914, p. 27-33; map) quoted Walcott's Mount Bosworth section to which he added Allan's Chancellor and Ottertail formations. He also reproduced Allan's geologic map of the rocks between Castle Mountain and Ross Lake.

In 1915 Walcott (1915, p. 162-233) published a comprehensive summary of the Cambrian events in the Cordilleran region. He gave hypotheses explaining the Cambrian—pre-Cambrian unconformity, the diastrophism during the period, the paleogeography, the climate, the source of the Cambrian faunas and the origin of the sediments, and indicated some of the unsolved problems pertaining to the Cambrian period in the Cordilleran region. The Cambrian sections, formations, and fossils of the Canadian area form an important part in these hypotheses. In this paper Walcott (1915, p. 178, 212) quoted the Mount Bosworth section and gave a faunal list of the species in the *Ogygopsis* shale member of the Stephen formation on Mount Stephen.

In 1916 Burling (1916a, p. 99-100; 1916b, p. 469-472) announced his discovery of the *Albertella* fauna, *in situ*, in the Cathedral formation in the Mount Bosworth and Castle Mountain sections, thus establishing the stratigraphic horizon of this important biota. He placed the Middle-Lower Cambrian boundary immediately above the *Olenellus* zone in the basal part of the Mount Whyte formation, thereby assigning most of the formation to the Middle Cambrian. In the latter paper Burling (1916b, p. 470) also summarized the Castle Mountain and Mount Bosworth sections but gave different thicknesses for the formations than those of Walcott.

In the same year Walcott (1916a, p. 222-228; 1916b, p. 303-457) described a number of new genera and species from the Cambrian strata of British Columbia but added no new sections nor information concerning the stratigraphy of the area.

During 1917, following Burling's discussion of the *Albertella* fauna and the Middle-Lower Cambrian boundary, Walcott published three papers which have caused much confusion in the stratigraphy, nomenclature, and paleontology of the area (Walcott, 1917 a, p. 1-8; 1917b, p. 9-59; 1917c, p. 61-114). In the first of these papers Walcott defined the new Ptarmigan formation and the Ross Lake shale which he doubtfully assigned as

a member of the Ptarmigan. He also listed the *Albertella* fauna which occurs only in the Ross Lake shale or at that horizon (Walcott, 1917a, p. 1-4).

In the next paper, published the same month, Walcott gave descriptions of the species comprising the *Albertella* fauna. Now, however, he combined into this fauna species from the Chetang limestone of the Robson Peak area, the Ptarmigan limestone from Ptarmigan Peak, the Gordon shale from Montana, and the Ross Lake shale in the vicinity of Kicking Horse Pass, British Columbia. Thus species from four areas (two of which are separated more than 600 miles) and from four formations were combined into one fauna and were implied to be of the same age. Later the same year Walcott (1917c, p. 61-114) described the fauna of the Mount Whyte formation. In this paper he lumped together specimens which occur throughout a thickness of 275 feet or more of strata and indicated their stratigraphic position as lower, middle, or upper within the formation. Such Middle Cambrian genera as *Kochaspis* were included in this "Lower Cambrian fauna," probably because of the error of the distribution of *Olenellus* which was thought to be in the lower, middle, and upper parts of the Mount Whyte, whereas this genus is restricted to the basal fourth of the formation.

In 1922 Burling (1922, p. 460-461) again suggested that the larger and upper part of the Mount Whyte should be transferred to the Middle Cambrian as the Middle-Lower Cambrian boundary probably is somewhere in the lower part of the Mount Whyte formation.

In 1923 Walcott (1923, fig. 24) correlated the formations in the Saskatchewan River area of Alberta with those in the combined Mount Bosworth and Field map area sections but gave no evidence for the correlations.

In 1927 Walcott (1927, p. 150-169; pl. 1) discussed the pre-Devonian sediments in the southern Canadian Rocky Mountains and concluded that five Paleozoic geosynclines ("troughs") could be recognized. The area studied in 1938 lies in the central eastern part of Walcott's Bow Trough.

The following year Resser assembled all Walcott's sections measured in Alberta and British Columbia, quoted the definitions of all the valid pre-Devonian formations used in the two provinces, and edited an unfinished manuscript left by Walcott. This paper (Walcott, 1927, p. 175-368), published posthumously, contains the first detailed sections of Castle Mountain and Ptarmigan Peak and will probably remain as the classic pioneer work on the early Paleozoic stratigraphy of the Canadian Rockies and as the starting point for all future work on these sediments and fossils.

In this paper, Walcott (1927, p. 187) summarized his years of exploration and reconnaissance in the Canadian Rockies by saying:

> "My study of it has been of the nature of a reconnaissance, made with the view of furnishing to the future areal and structural geologist some additional data on the succession of the pre-Devonian fossil faunas and faunules in the various sedimentary formations that collectively form one of the great pre-Devonian sections of the world."

In 1929 Warren (1929, p. 23-24) summarized the stratigraphic sequence along the 51st parallel. Warren used Walcott's nomenclature for the Cambrian formations in the area, including the invalid name, Fairview, for the Fort Mountain sandstone. Warren clearly stated the problem of the stratigraphic relationships of Allan's Chancellor shale in the Field map area to the Sherbrooke limestone on Mount Bosworth and gave Allan's latest opinion that "the lower beds of the Chancellor shale represent the Sherbrooke formation." Warren also summarized the evidence for drawing the Cambro-Ordovician boundary at the top of the Goodsir shale and proposed the name Fieldian for the Upper Cambrian series in Alberta and British Columbia.

In 1931 Raymond (1931, p. 97-116) published a structure section across the Canadian Rockies, part of which extends from Fossil Mountain to Mount Bosworth. In connection with the section Raymond gave a stratigraphic section, "after Walcott," in the vicinity of Lake Louise for which he combined the thicknesses of Walcott's formations in the Mount Bosworth and Ross Lake sections and the Sabine limestone of Schofield (1920, p. 76) from the Rocky Mountain Trench. Aside from such discrepancies, he omitted the Chancellor and Ottertail formations which occur in the area through which his section passes, southwest of Mount Bosworth.

The following year Resser and Bridge (1932, pl. 1) published a correlation chart of the Cambrian formations in the United States, in which they gave a composite section for the Canadian Rockies, using Walcott's formational names taken from his sections on Castle Mountain, Ptarmigan Peak, and Mount Bosworth.

In 1933 Resser (1933, p. 748-751) attempted to formulate a tentative standard Cambrian time and rock scale for North America. He erroneously placed the Ptarmigan in "British Columbia" above the Chisholm of Nevada; the Cathedral of British Columbia above the Ute of Utah; and the Eldon as older, or doubtfully equivalent in age to the Marjum of Utah.

In 1936 Grabau (1936, p. 42-64) comprehensively summarized much of Walcott's work on the Lower Cambrian in Alberta and British Columbia and discussed parts of the Ptarmigan Peak, Castle Mountain, Mount Stephen and Mount Bosworth, and Ross Lake sections. In his analysis of the Mount Whyte fauna Grabau (1936, p. 45-46) emphasized the fact

that *Kochaspis cecinna* (Walcott) is more typical of the Middle than of the Lower Cambrian.

In the same volume Grabau (1936, p. 219-233) discussed the Middle Cambrian rocks and fossils in the area. Because of the erroneous assignment of the *Albertella* fauna to the Ptarmigan formation he deduced that a disconformity is present between the Mount Whyte and Ptarmigan formations. The following year Grabau (1937, p. 251-254) quoted the Upper Cambrian part of Walcott's Mount Bosworth section and discussed the Upper-Middle Cambrian boundary on Castle Mountain and Mount Bosworth (Grabau, 1936, p. 246-247).

The last paper in which the Cambrian rocks of the area were mentioned was by Deiss (1939, p. 54, fig. 6), who correlated the Cathedral dolomite with part of the Gordon shale, the Stephen formation with the combined Damnation, Dearborn, Pagoda, Pentagon, and Steamboat formations, and the Eldon dolomite with the Devils Glen dolomite of northwestern Montana.

Since 1917 little if any field work has been done upon the stratigraphy of the Lower and Middle Cambrian formations in this part of Alberta and British Columbia, except in 1938 by the writer. The results of this work are given in the following pages.

CASTLE MOUNTAIN AREA, ALBERTA

GENERAL STATEMENT

The Castle Mountain section is one of the first mentioned in Walcott's reports on the Cambrian rocks of the Canadian Rockies (Walcott, 1908a, p. 2). Likewise, the Eldon, described from its type locality on Castle Mountain, was among the first named formations in the area. In Walcott's first Castle Mountain section, the strata were divided into five formations of which one, the St. Piran, was assigned to the Lower Cambrian, and four, Mount Whyte, Cathedral, Stephen, and Eldon, to the Middle Cambrian (Walcott, 1908a, p. 2). In spite of many references to the "Castle Mountain section" in the literature during the succeeding years a detailed section was not published until 1928 (Walcott, 1928, p. 274-277). However, in 1916 Burling (1916b, p. 470) gave a list of the Middle Cambrian formations, except the Eldon, on Castle Mountain and assigned to them different thicknesses than those given by Walcott. Burling also placed the Mount Whyte in the Middle Cambrian but correctly located the stratigraphic position of the *Albertella* zone in the Cathedral dolomite.

In his section published in 1928 Walcott assigned the dark gray and black limestones at the base of the Middle Cambrian to the Ptarmigan formation, and the shales, dolomites, and limestones above the Eldon to

the Upper Cambrian Arctomys and Bosworth formations. Unfortunately he stated that *Albertella* occurs in the upper 12 feet of the Ptarmigan limestone on Castle Mountain.

Because of these discrepancies in the literature and because of the strategic geographic and stratigraphic significance of the Lower and Middle Cambrian sediments in the vicinity of Castle Mountain, the section there had to be remeasured carefully and the exact horizons of newly collected faunas accurately located within the section.

Castle Mountain forms the southwest end of the structural, physiographic and stratigraphic province which extends northwest beyond Ptarmigan Peak (Fig. 1). This province is bounded on the northeast and southeast by Johnston Creek and the Johnston Creek fault, and on the south and southwest by the glacial drift and recent alluvium of Bow Valley. Within the area the rocks are Beltian shales, argillites, sandstones, and conglomerates overlain by Lower, Middle, and Upper Cambrian sandstones, limestones, dolomites, and shales. The Upper Cambrian is poorly represented within this province as only 150 feet to 1250 feet of these rocks have not been removed by erosion.

The strata are folded into an open, more or less symmetric syncline whose nearly horizontal axis trends N. 40° W. The beds on the east side of the syncline are locally dragged into steeper dips along the great overthrust which threw the Hector and locally the Corral Creek formations (Beltian) against Ordovician to Devonian limestones on the east. Johnston Creek follows the trace of this thrust, which sharply separates the Sawback Range (Fig. 1) from the unnamed group of peaks of which Castle Mountain is the most southwestern. Pleistocene glaciers have cut enormous elongated cirques along the axis of the syncline, and the peaks upon the long high-walled ridges on the southwest and northeast parts of the area are partly the result of this erosion. Thus the mountains in this province are strikingly different topographically from those in the Sawback Range to the east but are similar to those in the Bow Range to the west. Although slightly different topographically, Ptarmigan Peak and Redoubt Mountain are stratigraphically and structurally similar to Castle Mountain but are known locally as part of the Slate Mountains.

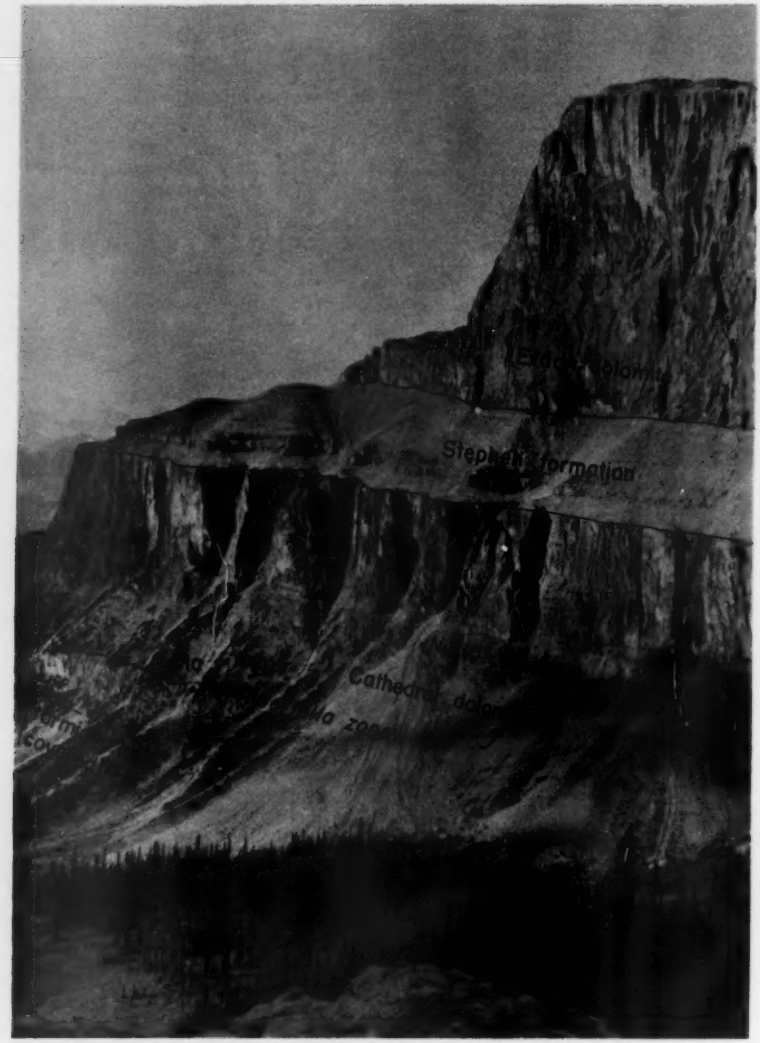
LOCATION OF AREA

Castle Mountain forms the east side of Bow Valley for a distance of more than 5 miles. Its sheer castellated walls rise several thousand feet above the valley floor (Pl. 1, fig. 2) between the towns of Castle Mountain and Eldon on the Canadian Pacific Railway, approximately 14 miles southeast of the town of Lake Louise and 18 miles northeast of Banff,

FIGURE 1. UPPER CIRQUE BETWEEN CASTLE MOUNTAIN AND HELENA RIDGE
B=Bosworth, A=Arctomys, P=Pika, and E=Eldon formations.

FIGURE 2. WEST SIDE OF CASTLE MOUNTAIN, BOW VALLEY IN FOREGROUND
E=Eldon, S=Stephen, C=Cathedral, Pt=Ptarmigan, M=Mount Whyte, and SP=St. Piran formations.

WEST SIDE, AND CIRQUE ON EAST SIDE OF CASTLE MOUNTAIN



SOUTHEAST SIDE OF CASTLE MOUNTAIN
Showing position of *Albertella*-zone in Cathedral dolomite.

Alberta. The mountain is near the western edge of Banff National Park, 6 to 9 miles northeast of the Continental Divide.

CASTLE MOUNTAIN SECTION

The Castle Mountain section was measured from the base upward, starting on the south side of Castle Mountain in the St. Piran sandstone. The traverse line of the section passes over the limestone cliffs of the Mount Whyte and Ptarmigan formations on the southwest side of the lowest cirque on Castle Mountain. The contact of the Stephen shale upon the Cathedral dolomite was traced to the southeast end of Helena Ridge where the entire Stephen formation (Pl. 3) is well exposed. The contact of the Eldon dolomite upon the upper Stephen limestone was traced to the northeast side of the middle cirque, and the section continued there to the top of the hill 9300 feet in elevation on Helena Ridge (Pl. 4), where the Upper Cambrian is slightly thicker than on Stewart Knob. The section was measured in secs. 8, 9, 16, 17, 20, and 21; T. 27 N., R. 14 W.[3]

UPPER CAMBRIAN

	Feet	Meters
Bosworth limestone		
Limestone: dark-gray, fine-grained, slightly fetid, thin-bedded. Upper beds finely oölitic, and contain drab-tan clay flakes. Occasional algal bed, *Girvanella?*, present. Limestone weathers drab-tan and dark-gray and forms sheer cliffs at top of "9300 Peak"	48	14.7
Limestone: black-gray, thin- and nodular-bedded in 2- to 8-foot zones, alternating with pale-gray reef-like beds or masses 2 to 4 feet thick. Upper surfaces hemispheric, 2 to 5 feet in diameter. Some beds petroliferous and marked with light-gray spots and narrow lines	39	12.0
Limestone: dull-gray, fine-grained, argillaceous, extremely platy. Weathers buff-gray and forms plates strewn over darker talus slope below	9	2.8
Limestone: pale-gray, oölitic, thick-bedded; interbedded with dark-gray, fine-grained, nodular-bedded limestone which contains clay flakes. Weathers dull-gray and forms cliffs	31	9.5
Limestone: dark-gray, fine-grained, argillaceous, nodular; contains lighter-gray clay flakes which give limestone mottled appearance; weathers dark-gray	5	1.5
Limestone: dark-gray, "reef"-like masses or beds; upper surface wavy; contains many tan clay flakes	5	1.5
Limestone: dull-gray, argillaceous, thin- or shaly-bedded. Interval largely covered	13	4.0
Dolomite: dull-gray, laminated, extremely thin- and uniform-bedded; weathers light-buff-gray and forms plates strewn on slope below. Upper 1 to 2 feet a "clinker" bed of orange-tan, pure limestone underlain by buff, argillaceous, thin- and flaggy-bedded limestone	20	6.1
Limestone: black-blue-gray, massive, fine-grained, algal; upper surfaces slightly hemispheric; weathers dark-gray	5–6	1.5–1.8

[3] The section numbers are based upon the Canadian system of numbering, which, unlike that used in the United States, begins in the lower right-hand corner of a township.

	Feet	Meters
Dolomite: dull-gray, nearly lithographic, thick- and thin-bedded (4 to ¼ inches); weathers white-buff-gray and forms plates	6–7	1.8–2.1
Limestone and dolomite: as underlying interval, but generally thicker-bedded, and more strikingly mottled black and light-gray	53	16.3
Limestone: black-gray and blue-black, fine-grained, massive, thick-bedded, mottled light- and darker-gray in some beds; others contain ovoid Algae, *Girvanella*, ⅛ to ⅜ inch long diameter. Limestone weathers dark-gray and forms cliffs	69	21.2
Limestone and dolomite: blue-gray, fine-grained, impure limestone in lower beds; and buff and drab-gray zones of hard, fine-grained, thick-bedded dolomite. All slightly banded. Upper 10 feet banded gray and buff-tan and underlain by 12 to 18 inches bright-orange-buff "clinker" layer of pure calcite. Upper 10 feet forms second or upper bench in cliffs here and on Stewart Knob	41	12.6
Shale: dull-gray on fresh fracture, platy to fissile; weathers pale-buff and covers slope of first bench in cliffs here and on Stewart Knob	2–3	0.6–0.9
Dolomite: dull-gray, fine-grained, hard, massive; weathers tan or drab-gray; forms cliffs and breaks down into angular fragments. Lower 4 to 5 feet thinner-bedded and finely banded pale-gray and buff on fresh fracture	36	11.1
Limestone: blue-gray, medium-grained, contains flakes and bands of limonitic tan clay and vaguely defined *Girvanella?*. Upper surface hemispheric and reef-like	1.5–2	0.5–0.6
Dolomite, limestone, and shale: alternate zones; green-, tan-, buff-, gray-, and drab-weathering, platy, shaly. Upper part limestone: tan-weathering, massive, hard, pale-buff-gray, dolomitic, medium-grained. Upper 2 feet of this interval dark-gray, slightly fissile, hard shale	43	13.2
Total thickness of Bosworth limestone	430	131.9

Note: Walcott may have drawn his Arctomys-Bosworth boundary lower in section, but this is the only natural boundary between the dominantly maroon shaly beds of the upper Arctomys and the limestone and dolomite of the Bosworth and is the only useful boundary for mapping.

Arctomys formation

Dolomite: alternate zones bright-buff-tan and maroon, 2 to 6 feet thick. Maroon beds finely arenaceous, and finely banded gray and lavender-maroon. Buff beds fine-grained, argillaceous. Many zones shaly-bedded, but no true shale in this interval	65	20.0
Limestone: pale-gray and buff-tan alternating in lower third. Tan-buff in upper two-thirds. One 1-foot zone of maroon platy shale in middle part of interval	30	9.2
Limestone and dolomite: dull-, dark-, and some buff-gray, all slightly fine-grained; buff-weathering. Upper beds weather dull-gray and finely banded on joint faces. Poorly preserved *Obolus?* sp. occur in drab-gray, argillaceous, fine-grained dolomite 8 feet below top	46	14.1
Shale: dull- and dark-maroon, becomes lavender-maroon and more calcareous up section. Finely wavy banded gray and maroon on fresh fracture. Weathers dark-lavender-maroon but generally covered with buff- and bright-tan talus from overlying beds	62	19.1
Shale: dark-maroon, dolomitic, finely arenaceous, hard, irregular, platy. Some beds dull-green and others buff; as much as 1 inch thick; rippled. Buff beds gray, argillitic dolomite on fresh fracture. Pseudomorphs of salt crystals as much as ¾ inch in		

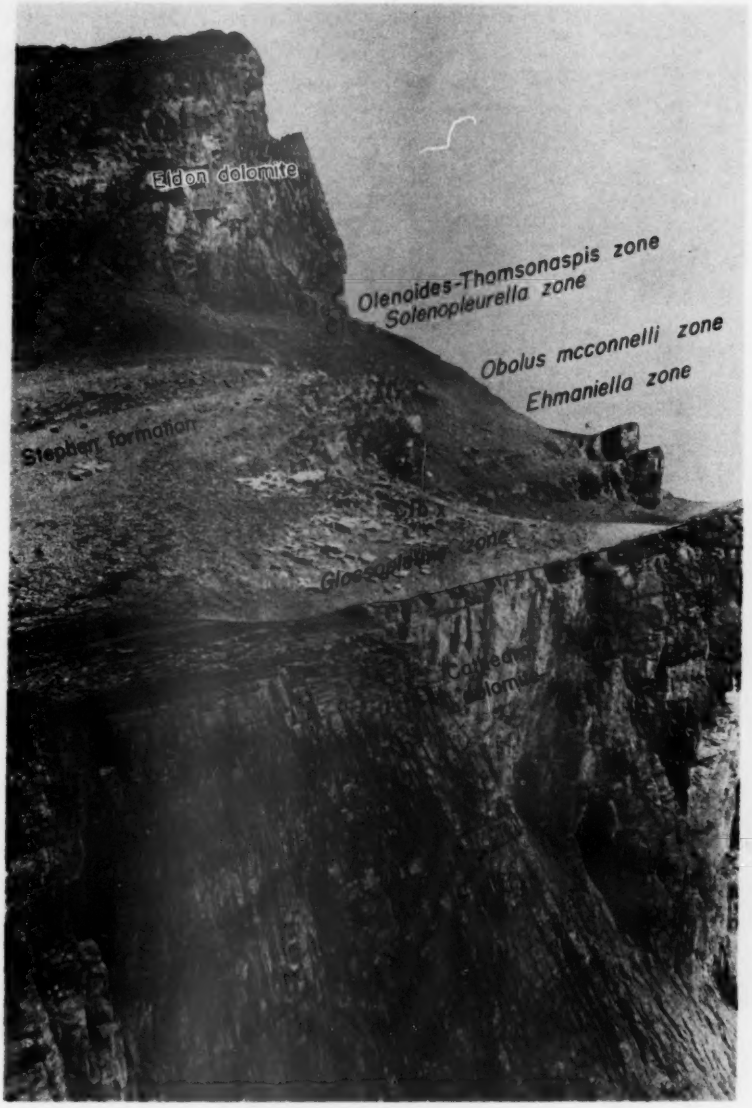
	Feet	Meters
length in some beds. Mud cracks present but not common. Weathers dull-maroon, but slopes usually covered with buff talus from overlying beds	51	15.7
Dolomite: dull-gray, fine-grained, irregular- (nodular) bedded (1 to 4 inches). Some beds flecked red on fresh fracture. Weathers bright-buff above gray smooth-bedded dolomite and below maroon shale	16	4.9
Total thickness of Arctomys formation	270	83.0
Total thickness of Walcott's Upper Cambrian	700	214.9
Pika formation (type locality)		
Dolomite: dark- and dull-gray, fine-grained, thick-bedded; weathers white or light-gray. Upper surface wavy or rounded. Forms light-gray cliffs on east side of Helena Ridge	24	7.4
Dolomite: darker-gray but weathers white-buff-gray as in underlying interval. All slightly banded light- and dark-gray	25	7.7
Interval covered	31	9.5
Dolomite: as in underlying interval but darker-gray, thicker-bedded, and more massive	106	32.6
Dolomite: as underlying interval but becoming thicker-bedded up section. Occasional beds mottled salt and pepper in middle of interval	43	13.2
Dolomite: dull-gray, uniformly thin-bedded (½ to 2 inches); contains varying amounts of buff arenaceous clay flakes. Weathers dull-gray	77	23.7
Dolomite: pale-gray, fine-grained, massive, hard, thick-bedded (6 to 18 inches). Weathers pale-buff and forms upper part of yellow-tan zone	10	3.1
Limestone: pale-gray, shaly; contains small ripple marks; weathers pale-gray	3	0.9
Dolomite: dull-, dark-, and white-gray; weathers bright-yellow-tan and to fine wavy bands on joint faces. Forms ledge and is prominent marker	6	1.8
Dolomite: as upper part of underlying interval. This interval largely covered	63.5	19.5
Dolomite: dull-gray and buff banded in lower part; becomes paler-gray, more vitreous, and thicker-bedded upward. Weathers gray and orange-tan and forms cliffs	30	9.2
Limestone: blue-gray, thin-bedded. Buff, arenaceous clay partings form wavy bands on joint faces	20	6.2
Limestone and shale: limestone thin-bedded, weathered gray, blue-gray, and buff. Occasional bed of intraformational conglomerate and oölitic limestone. Shale dull-brown-gray, argillaceous, fissile. Intercalated fossiliferous limestones contain numerous trilobite cranidia and pygidia. Fossil loc. C2a	18	5.5
Fossils: New genera and species.		
Limestone: one bed; blue-gray, oölitic; contains some intraformational conglomerate and bright-tan clay flakes	1–1.2	0.3
Limestone and covered interval: limestone thin-bedded	18	5.5
Limestone: one bed; blue-gray, oölitic; contains black and orange (limonitic) oölites and occasional pebbles of pale-gray, fine-grained limestone, and bright-tan clay flakes	0.7–0.8	0.21–0.25
Limestone: dark-gray, fine-grained, thin- and thick-bedded (1 to 5 inches); contains tan arenaceous clay flakes and partings. Interval largely covered	31	9.5
Limestone: pale-buff-gray and dark-blue-gray banded in wavy lines in lower part; thicker-bedded and less banded in upper part. Contains flakes and partings of tan arenaceous clay. Weathers buff and forms broken cliffs in slope above Eldon cliffs	10	3.1

	Feet	Meters
Interval covered: basal 2 feet dolomite; pale-gray, fine-grained, in 1- to 3-inch beds	25	7.7
Dolomite: pale-gray, thin- and flaggy-bedded. Largely covered	4.5	1.4
Dolomite: pale-gray, fine- to medium-grained, argillaceous, thin-bedded (½ to 12 inches); weathers banded 1/16 to ¼ inch on joint faces and to pale-buff-gray thin slabs. Indistinct mud cracks in some beds. Poorly developed edgewise conglomerate in one bed	3	0.9
Total thickness of Pika formation	550	168.95
Total thickness of Upper Cambrian	1250	383.85

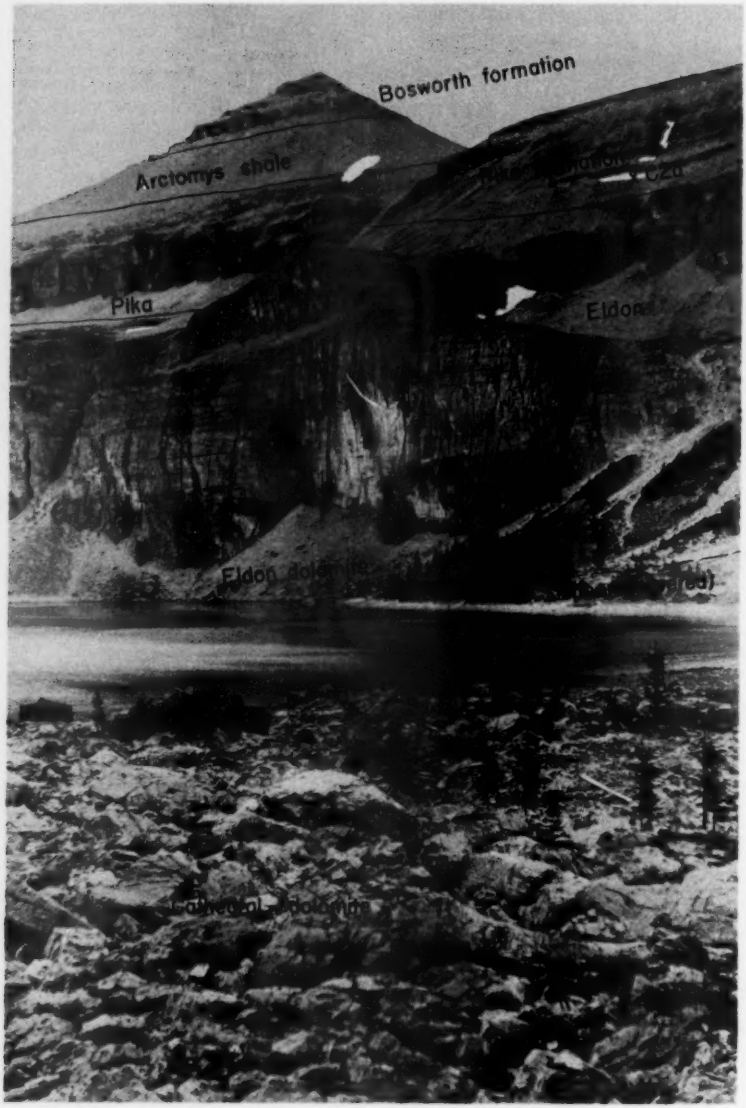
MIDDLE CAMBRIAN

Eldon dolomite (type locality)

Dolomite: dull-pale-gray, fine-grained, very slightly petroliferous, thin-bedded (2 to 4 inches). Forms covered slope above crest of Eldon dolomite cliffs	15	4.6
Dolomite: black- to dull-gray, medium crystalline, strongly fetid, thick-bedded or unbedded. Contains cavities lined with red hematite and amorphous oölites. Weathers black-gray and to vesicular surfaces	14	4.3
Dolomite: dull-, pinkish-, and tan-gray, fetid, more or less thin-bedded; weathers dark- or dull-gray and to rough surfaces	23	7.1
Dolomite: pale-blue-gray, finely crystalline, vitreous, thick-bedded; becomes darker-gray up section. Weathers dull-gray and forms cliffs	45	13.8
Note: May be 8- to 15-foot error between these two intervals.		
Dolomite: black-gray, petroliferous, rough weathered. Contains small (½ to 1 inch) irregular cavities which are occasionally lined with bright-crimson hematite	118	36.2
Dolomite: same as underlying but darker- (black) gray, thicker-bedded, and weathers to larger solution cavities. Here top of gray Eldon cliffs and of knoll above lake	127	39.1
Dolomite: dark-dull-gray, medium- to fine-grained, massive, thick- and thin-bedded, nearly black in upper part. Weathers dull-gray and to sharp points. Forms lower part of knoll between upper and lower Eldon cliffs northeast of lake	177	54.4
Dolomite: blue-black-gray, fine- and medium-grained, vitreous, thinner-bedded dolomite. Weathers dull-gray, to finely lined or mottled, rough, pitted surfaces	39	12.0
Note: Small fault may cause error of 16 or 20 feet in this part of section.		
Interval largely covered: underlain by pale- and black-gray, thin-bedded dolomite, which in cliffs weathers darker than underlying	27	8.3
Dolomite: darker-gray, medium crystalline, more regular-bedded (2 inches to 2 feet). Weathers irregularly dark- and light-gray, to irregular surfaces, and to fine bands or irregular white lines	37	11.4
Dolomite: same as underlying but slightly darker and more regularly bedded	66	20.3
Dolomite: white-gray, coarsely crystalline, thick-bedded. Similar to underlying beds	57	17.5
Dolomite: white-gray, hard, massive, thick-bedded, fine- to medium-grained. Weathers pale-gray. Upper beds finely banded and irregularly dark and light gray	30	9.2
Dolomite: white-gray pitted and banded; weathers white or light-gray	7	2.1



FAUNAL HORIZONS IN THE STEPHEN ON HELENA RIDGE,
CASTLE MOUNTAIN SECTION



WEST SIDE OF HELENA RIDGE, CASTLE MOUNTAIN SECTION
Showing type sections of Eldon and Pika formations.

	Feet	Meters
Dolomite: light-gray, medium crystalline, nearly unbedded; weathers dark-gray	20	6.2
Dolomite: white-gray, medium crystalline, deeply pitted in 2- to 3-foot bands in lower part. Weathers light-gray and to fine wavy bands. Two- to 4-foot zone of mottled or spotted dolomite. Dolomite passes laterally into dark-gray, more uniform-bedded dolomite	19	5.9
Dolomite: white-gray, medium crystalline, massive, hard; weathers light-gray and breaks down into angular blocks and fragments. In lower 6 feet, two (18 to 24 inches) bands of black-gray dolomite	25	7.7
Dolomite: black-gray, fine-medium-grained, extremely massive, in 2- to 6-inch beds (appears unbedded locally); slightly vitreous on fresh fracture, petroliferous; stylolites irregularly developed on many bedding surfaces. Algae? ⅛ to ¼ inch in diameter in lower beds. Weathers dull-gray and is usually covered by lichens which appear as black-gray weathering. Here top of Walcott's "basal 260-foot black interval"	81	24.9
Dolomite: one bed; sooty-gray, slightly vitreous, petroliferous, massive	5	1.5
Dolomite: dark- and tan-gray, thin-bedded (1 to 4 inches, average 1 inch), wavy and slightly nodular banded. Thicker-bedded upward. Weathers dull-gray and to sharp points and solution cavities (Pl. 5, fig. 2)	44	13.5
Dolomite: black-gray, fine-grained, uniformly thin-bedded. Bedding nodular and wavy. Weathers dark-gray	39	12.0
Note: Section traced northwest from top of the ridge to the base of the Eldon cliffs on northeast side of the lake and "tied in" on the thin-bedded, black-gray dolomite above the buff limestone of fossil loc. C1c. Accurate to within 5 or 10 feet.		
Total thickness of Eldon dolomite	1015	312.0
Stephen formation		
Limestone: dull-gray, fine-medium-grained, fossiliferous, thin-bedded (½ to 1½ inches), argillaceous; contains buff clay partings; weathers bright-buff. Fossil loc. C1e	5.7	1.7
Fossils: *Olenoides* cf. *serratus* (Rominger) *Solenopleurella* 2 n. sp. *Thomsonaspis* 2 n. sp.		
Limestone: dark-blue-gray, fine-grained, in beds ½ to 1½ inches thick; contains tan clay flakes and trilobites; weathers gray. Fossil loc. C1d	30	9.2
Fossils: *Ehmania?* sp. *Kootenia* sp. indet. *Solenopleurella* 3 n. sp. *Thomsonaspis* 2 n. sp. n. gen. of trilobites		
Shale: upper 5 to 6 feet drab- and green-gray, fissile; weathers gray, and contains *Obolus mcconnelli?* Walcott and small trilobite cranidia. Fossil loc. C1c top	5–6	1.5–1.8
Fossils: *Solenopleurella* sp. *Obolus mcconnelli?* Walcott		
Shale and limestone: green, green-gray, and some buff, calcareous shale; and some finely arenaceous, slightly micaceous, thin, platy, intercalated limestone. Weathers olive-green and buff and lighter than fresh fracture. Contains *Obolus* and trilobite cranidia and pygidia. Fossil loc. C1c in upper 40 feet	74	22.8
Fossils: *Ehmania?* sp. *Ehmaniella* cf. *walcotti* Resser *E.* 3 n. sp.		

	Feet	Meters
Elrathina? sp. *Solenopleurella?* sp. *Hyolithes* sp. *Obolus* sp.		
Intraformational conglomerate: one bed; gray, fine-grained limestone matrix. Pebbles pale-gray, fine-grained limestone as much as 5 inches in long diameter, average 1½ inches	0.4–0.5	0.12–0.15
Shale and limestone: same as underlying interval but more platy and calcareous. Contains lenticular slabs of rusty-tan-weathering limestone	30	9.2
Shale and limestone: same as underlying interval but paler-gray on fresh fracture; weather green-buff in some beds	13	4.0
Shale and limestone: shale buff and dull-gray, calcareous, thick-bedded; weathers buff and to irregular slabs. Occasional intercalated ½- to 1-inch beds of black-gray, argillaceous limestone; weathers to buff shaly slabs	18	5.5
Limestone: one bed; blue-gray, medium-grained; contains some intraformational conglomerate pebbles and grains of limonite	0.5	0.15
Shale: black-gray, fissile, hard, smooth; weathers very slightly lighter than color of fresh fracture	5	1.5
Limestone: 3 beds; dull-gray, medium-grained, hard; contains limonite grains, siliceous tan clay flakes, and occasional worm tubes	2.5	0.8
Shale and limestone: black fissile shale, and dark-gray, platy and nodular, thin-bedded limestone. Forms flat slope above cliff	5	1.5
Limestone: dark-gray, fine- to medium-grained, thick-bedded, lighter-gray and coarser-grained in lower part; contains orange limonite in middle part. Finer-grained and sooty-gray in upper beds. All beds weather dark-gray and to rough surfaces. Forms top of upper cliffs in lower part of Stephen	7	2.1
Limestone: dull-gray, thin-bedded, paler-gray upward. In upper third, occasional bed of intraformational conglomerate. Weathers pink-tan and finely banded on joint faces	14	4.3
Limestone: blue-gray, thin-bedded, platy, hard. Contains *Glossopleura* tails and free cheeks. Weathers gray. Top of *Glossopleura* zone and of fossil loc. C1b	10	3.1
Shale and limestone: clay shale; and intercalated, thin-bedded limestone; weathers yellow-buff	1.5	0.5
Limestone: dull-gray, finely-banded, slightly wavy-bedded; weathers reddish-tan and to sharp points and cavities; and forms top of lower cliffs in lower part of Stephen	1.5	0.5
Limestone: dull-gray, hard, massive but thin-bedded; becomes gray up section; weathers to pale-gray, thin slabs and fine bands	12.5	3.8
Limestone: dark-blue-gray, thin-bedded, handed with white calcite in some beds; weathers blue-gray and forms base of low cliffs	6.3	1.9
Limestone: blue-gray, fine-grained, thin-bedded (1 to 4 inches); weathers gray and to slabs. Occasional bed of intraformational conglomerate, and many trilobite fragments	31	9.5
Limestone: gray and blue-gray, fine-grained, platy (¼ to ¾ inch), and thin partings of tan clay. Occasional lenses of trilobite hash. Fauna similar to Damnation fauna of northwestern Montana. Base of fossil loc. C1b	6	1.8
FOSSILS: *Alokistocare?* sp. *"Amecephalus"* n. sp. *Chancia?* n. sp. *Glossopleura bosworthensis?* Resser *G. jordensis* Deiss *G. nitida* Resser n. gen. cf. *Inglefieldia* Poulsen *Kootenia* n. sp. *Zacanthoides* n. sp.		

	Feet	Meters
Limestone and shale: dull-gray, nodular, and platy, argillaceous, bright-buff-weathering limestone; and buff, clay shale partings	4–5	1.2–1.5
Total thickness of Stephen formation	285	87.3

Note: The section was traced from the cliffs on the southwest side of the amphitheater on Castle Mountain to the point approximately 8500 feet in elevation on the southeast end of Helena Ridge and was "tied in" on the contact of the upper Cathedral dolomite against the basal Stephen shale.

Cathedral dolomite

Dolomite: black-gray, fine-grained, thin-bedded (1 to 6 inches, average 2 inches); weathers dull-gray	6	1.8
Dolomite: pale-gray, coarsely crystalline, massive, almost unbedded; weathers tan-gray and forms top of upper Cathedral cliffs	60	18.4
Dolomite: pale-tan-gray, massive, thick-bedded or unbedded; weathers tan and forms sheer cliffs	80	24.6
Dolomite: black-gray, fine-grained, extremely thick-bedded; weathers pale-gray and in lower part to finely wavy banded joint surfaces. Alternate zones of pale-gray crystalline dolomite in lower part	42	12.9
Dolomite: dull-gray and tan-gray streaked dark-gray, coarsely crystalline in lower part; pale-buff-gray in upper part. Weathers tan	55	16.9
Dolomite: dark- and light-gray, medium crystalline, massive, thick-bedded; weathers tan on cliffs	30	9.2
Dolomite: steel- and black-gray mottled, medium-grained, hard, thick-bedded or almost unbedded	11	3.4
Dolomite: black- and dull-gray mottled, medium-grained, thin-bedded. Weathers buff-tan. Upper 6 to 10 feet forms bench in sheer cliffs. Stylolites well developed in this and lower intervals of the Cathedral between many beds	90	27.6
Dolomite: same as overlying interval but more uniformly dark- or black-gray and less mottled. Weathers gray, stained tan	48	14.8
Dolomite: dull-gray, fine-grained, thin-bedded (1 to 6 inches, average 2½ inches). Weathers deep-tan in contrast to gray above. Forms big bench in Cathedral dolomite cliffs. In basal 2 to 3 feet is 6- to 10-inch zone of black fissile fossiliferous shale (Ross Lake). This shale is fossil loc. C1a	50	15.4

FOSSILS: *Albertella helena* Walcott
A. robsonensis? Resser
A. similaris Resser
A. n. sp.
"Amecephalus" stator (Walcott)
Poliella? cf. *chilo* (Walcott)
Vanuxemella nortia Walcott
V. n. sp.
Hyolithes cercops Walcott
Obolus parvus Walcott
Paterina? wapta (Walcott)
Wimanella simplex Walcott

Dolomite: steel-gray and buff irregularly mottled, medium-grained, massive, hard, interbedded with tan and pale-gray, crystalline dolomite in beds ¼ to 3 feet thick. All weather buff-tan and form cliffs	64	19.7
Dolomite: white-gray stained pale-buff, coarsely crystalline, thick-bedded. Basal 2 to 3 feet "salt and pepper" mottled, gray and pale-tan. Upper fourth dull-gray, medium-grained, mottled drab-maroon on fresh fracture near weathered surfaces	71	21.8

	Feet	Meters
Dolomite: pink-white mottled black-gray, medium-crystalline, thinner-bedded than underlying. Weathers salmon-red. Top forms 2-foot undercut in cliffs	24	7.4
Dolomite: pale-buff-gray, some mottled dark- and pale-gray, coarse-grained, thick-bedded. Weathers buff-tan and to rounded sugary surfaces. Shallow solution cavities numerous	39	12.0
Total thickness of Cathedral dolomite	670	205.9

Note: The section was traced around the south end of the cliffs from the west to the east side of Castle Mountain on contact of the black-gray, upper Ptarmigan limestone against the pale-buff-gray, thick-bedded basal Cathedral dolomite.

Ptarmigan limestone

Limestone: black-gray, fine- to medium-grained, thin-bedded in upper half and nodular-bedded (1 to 3 inches) in lower half; and alternate zones of thin- and thick-bedded limestone. Small oölites in some beds, and limonitic clay flakes and partings in others	52	16.0
Limestone: black-gray, fine-grained, argillaceous, shaly, thin-bedded. Much sheared here and contains talc	4.5	1.4
Limestone: dull- to blue-gray, fine-grained, massive, thick- and thin-bedded. Contains oval algae and few clay flakes. Weathers dark-blue and to thin irregular beds	30	9.2
Limestone: dull-gray, fine-grained, finely banded; contains many tan arenaceous clay flakes and partings but no Algae	3.5	1.1
Limestone: pale-gray, fine-grained, thick-bedded; contains tan clay flakes in some beds and limonitic grains in others. Some beds contain small Algae (Pl. 5, fig. 1). Coarse oölite common in darker beds	9	2.8
Limestone: blue-gray, fine-grained, massive, thick-bedded or unbedded; *Girvanella?* as much as ½ inch in long diameter. Taken tentatively as basal Ptarmigan	6	1.8
Total thickness of Ptarmigan limestone	105	32.3
Total thickness of Middle Cambrian	2075	637.5

LOWER CAMBRIAN

Mount Whyte formation

Limestone and shale: blue-gray, coarsely oölitic, thick-bedded limestone, which contains trilobite fragments. Interbedded, dull-green-gray, medium-grained, platy- and irregular-bedded limestone, and buff-gray, finely arenaceous, chunky shale. Interval weathers to buff shaly slope below cliffs	53	16.3
Limestone: 2 beds; blue-gray, coarsely oölitic	2.3	0.7
Limestone, shale, and sandstone: blue-gray, coarsely oölitic, thick-bedded, and dull-gray, fine-grained, thinner-bedded limestone; finely banded, arenaceous shale; and thin-bedded, interbedded drab-green-gray micaceous sandstone. Worm trails on some sandstone beds	19	5.8
Shale, sandstone, and limestone: brown, maroon, and pale-gray-green shale in alternate zones 2 inches to 3 feet thick; and intercalated, coarse- and medium-grained sandstone in ½- to 6-inch beds. Some thin, impure limestone in upper part of interval	40	12.3
Sandstone: dull-gray, coarse-medium-grained; in 2 beds. Contains numerous limonite grains. Some cross-bedding; weathers gray-buff and to rounded surfaces	4.5	1.4

	Feet	Meters
Sandstone and shale: sandstone thin-bedded, contains much limonite in specks. Weathers rusty-buff. Shale dull-maroon to tan, argillaceous, fissile to chunky. Forms upper ⅔ of interval	11	3.4
Sandstone: tan-gray, coarse-grained, cross-bedded, one or 2 beds. Forms broken ledge in section here	1.7	0.5
Shale: black, fissile, very slightly micaceous	1	0.3
Sandstone: 2 beds; tan-gray, coarse-grained, banded	4.5	1.4
Shale and sandstone: black-gray, dull-maroon, and some olive-buff, fissile, very slightly micaceous shale, and some drab intercalated, thin-bedded sandstone. Weathers buff and forms partially covered slope	8	2.5
Total thickness of Mount Whyte formation	145	44.6

Note: The above part of the section measured below the cliffs at the southwest side of the southeast end of Castle Mountain. The section was traced southeast 3 miles on the contact of massive 6- to 8-foot quartzite at the top of St. Piran against the basal black-gray shale of the overlying Mount Whyte.

St. Piran sandstone

Quartzite and shale: upper 6 to 8 feet massive, thick-bedded; underlying 13 to 16 feet, shale and some intercalated quartzite. Basal part of interval same as underlying. Weathers buff-gray	72	22.2
Quartzite and shale: pale-gray, thinner-bedded quartzite and less shale than in underlying interval	77	23.7
Quartzite: same as underlying but more cross-bedding	88	27.1
Quartzite and shale: white, massive, vitreous, cross-bedded. Some intercalated black-gray shale in middle third. Quartzite weathers pale-buff-gray	87	26.8
Quartzite: white and pale-gray, medium- to coarse-grained, extremely vitreous, in ½- to 5-foot beds; weathers rusty-buff. Wave marks on some beds	140	43.1
Shale, sandstone, and quartzite: alternate beds of black-gray, arenaceous shale; gray, thin-bedded (¼ to 1 inch) sandstone which contains worm tubes; and white-gray, thick-bedded, cross-bedded quartzite. Ripple marked in upper part of interval	93	28.6

Note: The Lake Louise shale is not recognizable on Castle Mountain but may be represented by this 93-foot interval. However, at Ptarmigan Peak, 13 miles northwest, the St. Piran is 815 feet and the Lake Louise shale 60 feet in thickness, whereas, if this 93-foot interval is the southern extension of this shale at Castle Mountain, the St. Piran is only 464 feet thick. As the quartzites below the Mount White at Castle Mountain are similar to those in the typical St. Piran to the north, and, as the formation at Mount St. Piran is much thicker and contains similar intercalated shales (Walcott, 1928, p. 252 and 303), all the quartzitic strata below the Mount Whyte formation in Castle Mountain section are tentatively placed in the St. Piran formation.

Quartzite and shale: dull-gray, thin- and thick-bedded (as much as 18 inches) quartzite; and dark-gray, arenaceous shale	22	6.8
Quartzite: white, massive, partly unbedded. Forms top of massive white quartzite cliffs	94	28.9
Quartzite: pure-white, medium-fine-grained, thick-bedded; forms lower part of massive white cliffs	34	10.5
Quartzite and shale: alternate intervals as underlying, but more shale	55	16.9

	Feet	Meters
Quartzite and shale: alternate intervals, white-gray quartzite in 1/6- to 2-foot beds; and black, micaceous, chunky, very finely arenaceous to fissile shale in zones 2 to 4 feet thick	27	8.3
Quartzite: white-gray, thick- and thin-bedded; spotted with tiny grains of limonite; contains lenses 1 to 8 inches in thickness of black, hackley shale; weathers buff and is much jointed	25	7.7
Interval covered:	11	3.4
Quartzite and shale: white-gray, fine-grained, vitreous, thin- and thick-bedded (2 inches to 2 feet); occasionally cross-bedded, slightly jointed	60	18.5
Interval covered:	42	12.9
Quartzite and shale: white and buff-tan, medium- to coarse-grained, in beds 1 inch to 2 feet in thickness; contains specks of limonite and hematite; weathers gray. Overlying beds weather rusty-tan stained red in irregular areas. Quartzite vitreous on weathered joint faces, and stained with pure hematite. Some cross-bedding. In upper part, 1- to 2-inch zones (partings) of buff-gray, micaceous, arenaceous, fissile shale. Forms brown cliffs here. Beds weather to rounded smooth surfaces. *Note:* The section begins at a point approximately 6200 feet in elevation, N. 20° E. of Storm Mountain, and N. 32° E. of Windimere road bridge over Bow River.	58	17.8
Total thickness of St. Piran sandstone	985	303.2
Total measured thickness of Lower Cambrian	1130	347.8
Total thickness of Castle Mountain section	4455	1369.15

COMPARISON OF ORIGINAL AND EMENDED CASTLE MOUNTAIN SECTIONS

In the original Castle Mountain section (Fig. 2) Walcott (1928, p. 274-277) estimated considerably greater thicknesses for nearly all the Cambrian formations than were obtained by carefully checked measurements. This is especially true of the Mount Whyte and Ptarmigan formations to which he assigned a combined thickness of 520 feet, whereas they possess a maximum thickness of only 250 feet. On the other hand, Walcott began his section on the southwest end of the mountain where only 55 feet of St. Piran sandstone is exposed. Several miles to the northwest on the same side of the mountain 985 feet of beds is exposed continuously beneath the Mount Whyte-St. Piran contact upon which this part of the section was accurately traced and "tied in" with that at the southwest end of the mountain. Consequently, the Castle Mountain section includes 985 feet of St. Piran instead of 55 feet as shown in the original section.

Fossils are rare and poorly preserved in the basal beds of the Middle Cambrian (Ptarmigan formation), and on Castle Mountain the sheer walls and steep slopes into which the exposures of the Ptarmigan and Mount Whyte strata are carved offer little opportunity of either finding fossils or obtaining good specimens. Consequently, the boundary between the Middle and Lower Cambrian is not accurately established in this

section but is inferred to be at the base of the black-gray Ptarmigan limestone at the horizon established upon evidence from fossils on Ptarmigan Peak 15 miles to the northwest (Fig. 1).

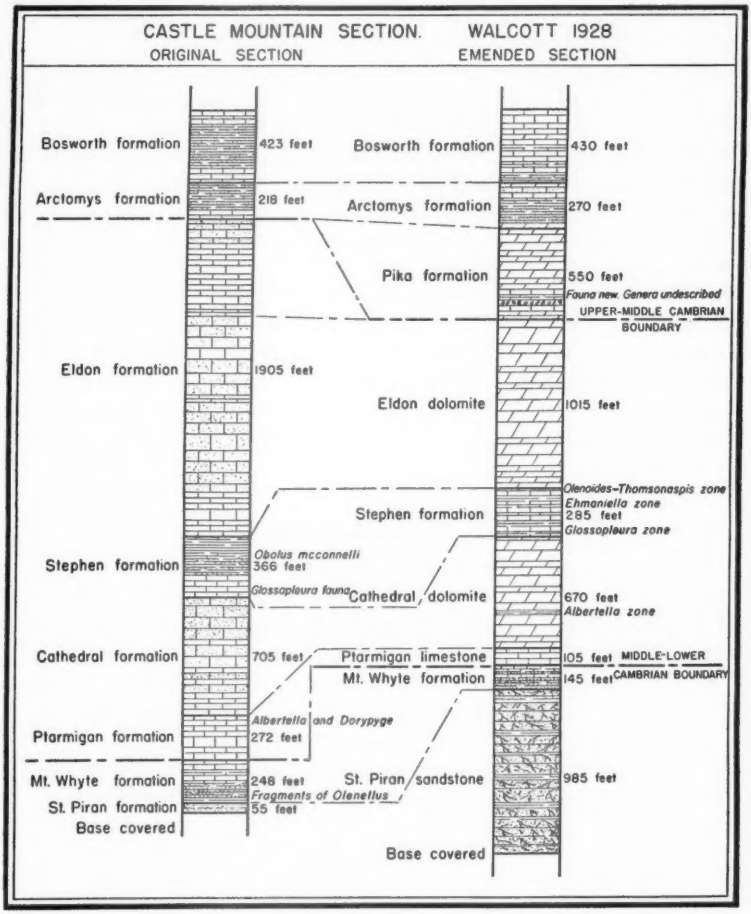


FIGURE 2.—*Comparison of original and emended Castle Mountain sections*

In the original Castle Mountain section, Walcott included in the Eldon formation all the rocks between the top of the Stephen limestones and the base of the "Arctomys" maroon and buff sandy shales. The lower

1015 feet of massive gray Eldon dolomite comprises a readily mappable formation and is different lithologically from the overlying 550 feet of shales and thin-bedded dolomites and limestones below the "Arctomys," which contains an easily recognizable fauna of upper Cambrian age and is also a natural readily mappable unit. Consequently, the upper 550 feet of the Eldon is a separate formation, the Pika.

The *Albertella* fauna, correctly located by Burling (1916b, p. 470), is well represented, but poorly preserved, on Castle Mountain where it occurs in a black fissile soft shale in the Cathedral dolomite approximately 200 feet above the base. However, fossils could not be found in the upper 12½ feet of the Ptarmigan limestone in which Walcott (1928, p. 276) stated that he found *Albertella*.

The application of the name Arctomys formation to the shales, sandstones, and dolomites between the Pika and Bosworth formations is of doubtful value but may be correct. The correlation of these rocks with those of the Arctomys in its type section 65 miles to the northwest is based upon the lithologic similarity and stratigraphic position of these rocks in the two areas but not upon fossils. Only *Obolus* sp. and *Lingulella isse* (Walcott) are known from the lower part of the formation near Glacier Lake (Walcott, 1928, p. 374).

PTARMIGAN PEAK—REDOUBT MOUNTAIN AREA, ALBERTA

GENERAL STATEMENT

The rocks which form Ptarmigan Peak and the surrounding area are folded into an open syncline with Ptarmigan Peak at the center. The rocks on Ptarmigan Peak are nearly completely exposed and contain many well-preserved fossils. As noted in the discussion of the Castle Mountain area, similar strata in this area are also thrust over Devonian limestones along the northern continuation of the Johnston Creek fault and are part of the same structural and stratigraphic province. This and the fact that the strata are continuous between these mountains cause the Ptarmigan Peak section to serve as an excellent check upon the stratigraphic and faunal sequence in the Castle Mountain section.

In 1917 Walcott (1917a, p. 1-4) published the original difinition of the Ptarmigan formation and its type section on Ptarmigan Peak. In that section he drew the boundary of the formation (correctly) at the base of a 62-foot shaly bluish-gray limestone and at least that distance below the horizon of *Kochaspis chares* (Walcott). In the same paper he stated:

> "It may be an error to include the Ross Lake shale with its *Albertella* fauna in the Ptarmigan formation. That is one of the problems for the future worker in this field to determine."

However, the previous year Burling (1916b, p. 470) had accurately located the *Albertella* shale (Ross Lake shale of Walcott) 200 feet and 375 feet above the base of the Cathedral formation on Castle Mountain and Mount Bosworth.

In 1928 Resser posthumously published Walcott's paper in which the Ptarmigan formation and section were again given, but the lower boundary of the Ptarmigan was placed 90 feet higher than in 1917, and the rocks in this thickness were assigned to the Mount Whyte and therefore to the Lower Cambrian (Walcott, 1928, p. 278-279).

The original Ptarmigan section (Walcott, 1917a, p. 2-5) was incomplete because the Mount Whyte, St. Piran, and Lake Louise were missing. On the other hand, the complete section from the pre-Cambrian Hector formation to the top of the Cambrian rocks in the area was given in 1928 (Walcott, 1928, p. 278-281), but an explanation was not made of the omission in the earlier section. However, before 1917 Walcott (1910, p. 319, 330, 331; 1916, p. 328) described several trilobites from the lower part of the Mount Whyte formation in Ptarmigan Pass. In this connection Resser (personal communication) states that:

> "Walcott measured the Ptarmigan Peak section in 1916. The difference between the 1917 and 1928 sections is simply that the latter is more complete because in the former he was intending merely to define the new formation." (Ptarmigan)

The abnormal thickness given for the Cathedral (2100 feet) also needed explanation, especially as the formation was reported to be only 1157 feet on Castle Mountain and 775 feet on Mount Bosworth, 15 and 10 miles to the southwest and west respectively (Burling, 1916b, p. 470).

Finally, the Ptarmigan Peak area was reported to contain one of the best exposures of the Cambrian-Beltian contact in this part of the Canadian Rockies and was the area in which Walcott (1910a, p. 428-429) measured the type sections of the pre-Cambrian Hector and Corral Creek formations.

In light of these facts, careful remeasurement of the Ptarmigan Peak section became imperative as one of the first steps in obtaining data for the revision of the Lower and Middle Cambrian stratigraphy in this part of the Canadian Cordillera.

LOCATION OF AREA

Ptarmigan Peak lies northeast of the head of Corral Creek, approximately 6 miles N. 47° E. of the town of Lake Louise, 15 miles N. 30° W. of Castle Mountain, and 10 miles N. 70° E. of Ross Lake (Fig. 1). The intersection of the parallel 51° 35′ and the meridian 116° 05′ is half a mile north of the mountain top. Ptarmigan Peak also lies in Banff National Park, in the province of Alberta.

PTARMIGAN PEAK SECTION

The Hector shale (estimated 700 feet) exposed below the Fort Mountain sandstone in this section exhibits many characters of Cambrian rocks in other areas, such as intraformational conglomerate beds with gray-green and maroon lithographic limestone pebbles, unmetamorphosed limestone and shale, and sandstones which contain quartz pebbles similar to those in the overlying Lower Cambrian sandstones. A disconformity is not apparent between the Lower Cambrian conglomerate and the Hector shale in this section. The Fort Mountain-Hector contact is transitional, the drab-green-gray Hector being overlain by a pebbly sandstone, in turn overlain by similar shale, which in turn is overlain by more (main mass) of the pebbly sandstone. The Cambrian—pre-Cambrian contact is arbitrarily taken at the base of the lowest pebbly sandstone conglomerate. The section was measured in the N. ½ secs. 17 and 18, SE. ¼ sec. 19, and SW. ¼ sec. 20; T. 29 N., R. 15 W.

	Feet	Meters
UPPER CAMBRIAN		
Pika formation		
Dolomite: dull-gray, paler-gray upward, and more vitreous. Some banding, weathers gray and orange-tan and forms cliffs	35	10.8
Limestone: blue-gray, thick- and thin-bedded; contains irregular bands and partings of buff-tan clay	25	7.7
Limestone and shale: blue-gray and yellow-buff, fossiliferous, thin-bedded limestone intercalated in dull-brown-gray, argillaceous, fissile shale. Weathers to shaly slope across top of saddle on peak top. Contains extremely numerous trilobites. Fossil loc. C4b	15–20	4.6–6.1
FOSSILS: New genera and species.		
Limestone and conglomerate: pale-buff and dark-blue-gray banded, fine-grained limestone in 1- to 5-inch beds. Some oölite and intraformational conglomerate in dark-blue, medium crystalline beds	40	12.3
Dolomite: dull-gray, fine- to medium-grained, in 1- to 8-inch beds; weathers banded pale-buff and gray and forms slabs. Forms buff bands above gray Eldon cliffs	30	9.2
Total thickness of Pika limestone on Ptarmigan Peak (estimated)	150	46.1
Note: Above detail taken from Castle Mountain section which is closely similar to that on upper part of Ptarmigan Peak. Thicknesses of above units are estimated.		
Total thickness of Upper Cambrian	150	46.1
MIDDLE CAMBRIAN		
Eldon dolomite		
Dolomite: dull-gray, medium-grained, hard, thick- and thin-bedded. Weathers dull-gray. Lower bed 5 feet thick. Upper beds 4 to 10 inches thick. Forms top of Eldon here	15	4.6
Dolomite: as underlying. Weathers gray and forms rounded ledges	57	17.5
Dolomite: pale- and some dark-gray. Weathers dull-pale-gray and to vesicular cavities	59	18.1

	Feet	Meters
Dolomite: black- and light-gray in alternating bands. More white-gray, medium-grained, thinner-bedded than in underlying interval. Weathers light- and dark-gray in irregular bands	53	16.3
Dolomite: dull-, tan-, and some dark-gray, medium- and coarse-grained, hard. Weathers light-gray as underlying	50	15.4
Dolomite: white-gray, medium crystalline, thick-bedded, finely and irregularly banded. Weathers light-gray	106	32.6
Dolomite: light- and dark-gray in alternating, irregular intervals 12 to 20 feet thick, medium- and coarse-grained, and some fine-grained. Some dark beds mottled white in tiny irregular spots and bands	55	16.9
Dolomite: light- and dull-gray, medium- and coarse-grained, weathers light-gray stained buff above darker-gray interval	59	18.1
Dolomite: black- and light-gray, medium-grained, slightly petroliferous, in 2- to 12-foot beds. Contains bands or veinlets of white, coarse-grained dolomite and irregular bands and blebs of tan weathering dolomite. Weathers dark-gray mottled light-gray	57	17.5
Dolomite: light- and dark-gray, coarse-grained, massive, thick-bedded or unbedded. Some beds dark-gray mottled white in small spots; other beds light- or tannish-gray mottled irregularly dark-gray. Occasional bands ⅛ to ¼ inch thick of coarser-grained dolomite. Dolomite weathers gray and buff-gray	53	16.3
Dolomite: dark-gray, fine- to medium-grained, thin- and platy-bedded. Weathers dark-gray and forms bench above lower sheer Eldon cliffs	90	27.7
Dolomite: as underlying. Some dark-gray in 5- to 12-foot bands in upper ⅔ which alternate with light-gray bands	172	53.0
Dolomite: white and some spots of dull-gray, extremely coarse-grained and thick-bedded. Weathers light-gray as underlying	78	24.0
Dolomite: light-gray, coarsely crystalline, massive, in 1- to 4-foot beds; weathers light-gray	156	48.0
Dolomite: dull-tannish-gray, medium crystalline, in 2- to 5-inch beds, slightly vitreous, banded light- and dark-gray. Many ½- to 2-mm. white calcite veinlets. Some gray, finely oölitic in lower part. Forms top of basal black zone	87	26.8
Dolomite: dark-gray, crystalline, thin- and platy-bedded. Forms base of dark-gray band in Eldon cliff	83	25.5
Total thickness of Eldon dolomite	1230	378.3

Stephen formation

Limestone: dull-gray in lower, sooty-black in upper part, fine-grained, extremely argillaceous, and shaly-bedded. Occasional 8- to 18-inch zone of bright-tan weathering, massive, thin-bedded limestone. Interval weathers dark-gray and forms top of Stephen slope below sheer dolomite cliffs of Eldon. Usually covered by dolomite talus	20	6.1
Shale and limestone: dull-green-gray, grayer upward, argillaceous, and some finely arenaceous, platy, and a little fissile shale; and a few beds of dull- and dark-gray, medium crystalline limestone. Weathers dull-green-gray, and in upper part gray, and forms smooth slopes	65	20.0
Shale and limestone: bright-green and tan, argillaceous, calcareous, and finely arenaceous, platy shale, and some intercalated orange-tan limestone which weathers to angular fragments. Interval weathers bright-buff-tan and forms slope	23	7.1
Limestone and conglomerate: one bed; tan-gray, medium-grained limestone matrix and tan-gray, finer-grained pebbles. Upper surface of bed hemispheric and composed of ½- to 2-inch layer of finely banded limestone which contains glauconitic clay, and rare trilobite cranidia. Weathers deep-tan	1–1.3	0.3–0.4

	Feet	Meters
Shale and limestone: tan-gray, calcareous, argillaceous shale, and dull-gray, fine-grained, slightly argillaceous limestone in ¼- to ¾-inch beds. Limestone and shale interbanded to form massive ledges which are wavy-bedded. Weathers tan-gray with dull-green areas in upper part. Near top a 3-inch bed of intraformational conglomerate. Pebbles tan-gray, fine-grained limestone coated with dark-green clay shale	47	14.5
Shale and limestone: dull-brown-gray, fissile, argillaceous shale and intercalated dull-gray, argillaceous, fine-grained, nodular, and lenticular limestones. Interval weathers buff-tan and forms slope. Occasional intercalated 1- to 4-inch beds of blue-gray limestone. Shale much sheared and slickensided	30	9.2
Limestone: one or 2 beds; dark-blue-gray, coarsely to medium crystalline, hard; contains flakes and blebs of tan clay and irregular spots of bright-orange-tan limonite	0.5–1	0.15–0.3
Shale and limestone: dull-gray fissile shale and a few dark-gray, intercalated, 1- to 4-inch beds of limestone. Interval weathers gray and is nearly everywhere covered with talus	8.5	2.6
Limestone: sooty-black, fine-grained, slightly argillaceous, lumpy, thin- and irregular-bedded (2 to 4 inches). Contains flakes and blebs of tan arenaceous clay, limonitic spots, and small pyrite crystals. Weathers dark-gray and forms top of cliffs	5.2	1.6
Limestone: dull-gray, medium-grained, finely banded, shaly; much sheared here. Weathers dull-pale-gray	5	1.5
Limestone: black-gray, fine-grained, in ¼- to 6-inch (average 1-inch) beds. Some blue-gray, finely crystalline limestone and many buff-gray clay partings. Weathers dark-gray and forms bulk of Stephen cliffs. Occasional *Glossopleura* still present	79	24.3
Limestone and shale: black-gray and pale-gray, fine-grained, platy, in ¼- to 1-inch beds. Contains many clay shale partings and flakes and numerous pygidia and cranidia of several species of *Glossopleura*, one of *Alokistocare* and *Kootenia*, and several small trilobite genera. Fossil loc. C4a. Limestone weathers to black and tan plates. Several intercalated beds of intraformational conglomerate and occasional lenticular bed of trilobite hash	19	5.8
FOSSILS: *Amecephalus?* 2 n. sp. *Chancia* 2 n. sp. n. gen. cf. *Glossocoryphus* Deiss *Glossopleura boccar* (Walcott) *G.* cf. *boccar* (Walcott) *G. fordensis?* Deiss *G. nitida* Resser *G.* cf. *walcotti* Poulsen *G.* 2 n. sp. n. gen. cf. *Strotocephalus* Resser *Zacanthoides* sp. indet.		
Shale and limestone: tan-gray shale and dull-gray, platy and shaly limestone. Weathers buff-tan and forms basal bench of Stephen. Lower part of *Glossopleura* zone. Fossil loc. C4a at base	11	3.4
Total thickness of Stephen formation	315	96.8

Cathedral dolomite

Dolomite: dark-steel-gray, fine- to medium-grained, slightly vitreous in irregular 1- to 2½-inch beds. Contains flakes of darker-tan-gray clay; weathers bright tan on joint faces and to platy beds. This is same thin-bedded dolomite that forms top of Cathedral formation in Castle Mountain section	4	1.2
Dolomite: dull-gray and some mottled dark-gray, medium-coarse-grained, thick-bedded, hard. Weathers tan-gray and some red-tan and to rough surfaces. Contains blebs and flakes of gray		

	Feet	Meters
and tan clay. Occasional zones of massive but platy-bedded dolomite	52	16.0
Dolomite: dark- and light-tan-gray, mottled pale-gray, medium-coarse-grained, in 1- to 2-foot beds; some platy 2- to 6-inch beds. Weathers bright-buff-tan-gray and forms rough surfaces	52	16.0
Limestone: black, as underlying, but more massive and thicker-bedded. Forms dark-gray cliffs below buff dolomite crags and talus	53	16.3
Limestone: black, fine-grained, in lower and upper thirds; finely banded pale-gray in middle third. Upper beds contain trilobite fragments. Weathers dark-gray and forms cliffs	60	18.5
Dolomite: dark- and light-gray mottled, medium-coarse-grained in 18- to 24-inch beds. In upper fourth average 2-inch beds of flaggy, pale-gray, banded, coarse-grained dolomite. Contains much tan limonite on joint faces. Stringers, or lenticular, irregular beds of black limestone locally developed throughout interval	72	22.1
Dolomite and limestone: dull-steel-gray, medium-grained, hard, slightly vitreous, thick- and thin-bedded dolomite (2 to 10 inches); dark- and light-tan and gray mottled dolomite; and in middle, 2- to 4-foot zone of black, fine-grained, thick-bedded limestone which contains *Girvanella*. Lower 8 feet dull-gray, sugary, calcareous dolomite. Interval weathers tan-gray and forms 2- to 4-inch slabs	57	17.5
Dolomite and limestone: black, fine-grained, argillaceous, thin-bedded limestone; contains *Girvanella* as much as 1 inch in diameter and indeterminable trilobite fragments. All much weathered. Limestone grades abruptly, laterally, to light-gray, coarse-grained dolomite. In a few tan-gray dolomite beds are oval areas which may represent dolomitized *Girvanella*	26	8.0
Dolomite: white-gray, becoming darker upward, medium-grained, thin-bedded (2 to 5 inches; average 2½ inches). Weathers buff-tan and forms bench	81	24.9
Dolomite: as underlying, but upper middle 10 feet mottled dark- and light-gray. Weathers pale-buff-gray	58	17.8
Dolomite: light-gray, extremely coarsely crystalline and thick-bedded. Alternate beds 4 to 8 feet thick of pale- and light-buff-gray dolomite. Occasional thin (8 to 14 inches) bed of dull-tan, finer-grained dolomite	50	15.4
Dolomite: dark-gray and thin-bedded in lower 8 to 12 feet. Upper part pale-gray, medium- and coarse-grained, in ⅔- to 5-foot beds, hard; weathers pale-buff-gray and forms large talus blocks	55	16.9
Total thickness of Cathedral dolomite	620	190.6

Ptarmigan limestone (type locality)

Limestone: pale-gray, slightly lithographic, massive, hard; weathers paler-gray than underlying and forms sharp contact with basal buff dolomite of Cathedral formation	3	.9
Limestone: as underlying. Contains extremely rare long-spined small much weathered pygidia. Fossil loc. C3f. This is Walcott's *Neolenus constans* horizon	23	7.1
Fossils: n. gen. cf. *Kootenia* Walcott		
Limestone: black-lead-gray, fine-grained, shaly-bedded in lower part and thin-bedded (1 to 6 inches) in upper part. Weathers black-gray above dull-gray	55	16.9
Limestone: as underlying. Forms top of massive sheer cliffs on northwest side of Redoubt Mountain and south side of Ptarmigan Peak	15	4.6
Limestone: black-gray, finely crystalline, slightly petroliferous, thick- and thin-bedded; contains many flakes of gray clay and		

	Feet	Meters
stringers and blebs of white calcite. Weathers dark-gray and to platy beds 1 to 2 inches thick	93	28.6
Limestone: dove-gray, fine-grained, some lithographic, thick-bedded; contains tiny white calcite blebs; weathers light-gray. Some poorly developed lenses of edgewise conglomerate occur in upper part	8	2.5
Limestone: dark- and drab-tan-gray, fine-grained (sugary), thick-bedded (½ to 2 feet), massive, hard. Some drab beds more petroliferous. Upper beds contain tiny veinlets and blebs of white calcite which weather in relief. Limestone weathers dark-gray and to sharp points. Trilobite fragments in some beds	39	12.0
Limestone: dark-gray, fine to medium crystalline in alternate zones of thin platy and thick massive beds. In upper half, thicker beds and poorly developed coarse and fine oölite. Limestone weathers dark-gray and to sharp angular fragments	50	15.4
Limestone: alternate zones of pale-dove-gray, slightly lithographic, finely banded; weathers light-gray and bands weather in relief; and dark-gray, fine-grained, thick-bedded limestone which contains stringers and blebs of white calcite. Near base, one or 2 beds of breccia of white-gray angular limestone fragments in dark-gray matrix. Interval weathers lighter-gray than overlying or underlying limestones	9	2.8
Limestone: as underlying, but more thick-bedded. Some black-gray, finely crystalline, smooth weathered beds in upper part	19	6.2
Limestone: as underlying. Some beds contain blebs and stringers of pale-gray calcite; others of gray clay. Weathers as underlying	33	10.1
Limestone: sooty-gray, massive, thick-bedded, slightly petroliferous; weathers dull-dark-gray	42	12.9
Limestone: black-lead-gray, fine-grained, argillaceous, nodular- and thin-bedded, massive. Contains flakes and blebs of buff and gray clay. Weathers dark-gray and to sharp points and cavities and forms cliffs. Some beds very slightly petroliferous	13	4.0
Limestone: dull-gray, fine-grained, argillaceous, thin- (platy) bedded; contains tan and gray clay partings, flakes, and nodules. Weathers dark-gray and forms part of dark cliffs. In upper middle, rare cranidia of *Kochaspis* cf. *cecinna* (Walcott). Fossil loc. C3e	23	7.1
Fossils: *Kochaspis* cf. *cecinna* (Walcott) *K.* cf. *gogensis* (Walcott) *Micromitra?* sp. indet.		
Limestone: dull- and lead-gray, fine-grained, extremely massive but thin-bedded, banded with tan clay and contains many tan and buff clay flakes, blebs, and nodules; weathers dull-gray. At lower third, an 8- to 10-inch zone of nodular, gray, fine-grained, fossiliferous limestone	35	10.7
Fossils: *Kochaspis?* n. sp. *Hyolithes* sp. *Lingulella?* sp. *Nisusia* sp.		
Total thickness of Ptarmigan limestone	460	141.8
Total thickness of Middle Cambrian	2625	807.5

LOWER CAMBRIAN

Mount Whyte formation (type locality)

Shale and limestone: gray, fissile, finely micaceous and arenaceous shale and little interbedded nodular limestone. A 2-foot bed of gray, fine-grained limestone in middle contains tan clay flakes and nodules. Interval weathers buff-gray and forms slope	13	4.0

	Feet	Meters
Limestone: dull-gray, platy-bedded; contains tan clay bands and partings. Some oölite in occasional thick bed	9	2.8
Interval covered: underlain by thin-bedded limestone and shale. Forms top of oölite zone and of buff slope below tan, platy weathering beds	6	1.8
Limestone: dull-gray, fine- to medium-grained, massive, thick- and thin-bedded; and much blue-gray oölitic and pisolitic lime stone	9.5	2.9
Limestone and shale: alternate zones of blue-gray oölitic limestone 1 to 2 feet thick, and buff-gray and tan- weathering zones of platy-bedded, gray limestone, and tan shale partings. Tan siliceous clay flakes and bands in some beds of blue-gray oölitic limestone, and worm tubes on shaly limestones	70	21.5
Shale, limestone, and conglomerate: dull-green-gray, micaceous, and finely arenaceous to soft shale; and interbedded green- and pale-gray, platy, lenticular, and occasionally finely arenaceous limestone. Worm tubes numerous on some limestones. In upper 2 feet, several 1- to 2-inch beds of blue-gray intraformational conglomerate. Fossil loc. C3c top	11	3.4
Limestone: blue-gray, coarsely oölitic, thick-bedded. At base, 1½- to 3-inch bed of dull-gray, finely oölitic, fossiliferous limestone which contains a faunule of dwarfed trilobites. Fossil loc. C3c	4	1.2
Fossils: n. gen. cf. *Amecephalina* Poulsen *Anatagmus?* sp. indet. *Ingleficldia? stephenensis* (Resser) *Kochiella crito?* (Walcott) *Plagiura cercops* (Walcott) *P.* n. sp. *Plagiurella cleadas* (Walcott) *P.* n. sp. *Syspacephalus?* sp.		
Limestone: blue-gray, fine-grained, contains tan clay partings	3.5	1.1
Limestone: dark-blue-gray, oölitic, thick-bedded. Oölites 1 mm. or less at base and increase almost uniformly in size upward to pisolites 7 to 8 mm. in diameter	1.5	0.5
Limestone, shale, and conglomerate: pale-gray, medium-grained, platy, finely banded limestone and interbedded buff-gray and green fissile shale partings. One 3- to 4-inch bed of intraformational conglomerate in upper 10 inches. Limestone weathers dull-gray banded tan and to sugary surfaces which simulate sandstone. Worm tubes on some beds in lower part of interval	11	3.4
Limestone: dark-blue-gray, oölitic; in 3 beds, separated by shaly, finely-grained limestone, and gray shale. Oölites 1 to 1½ mm. in diameter. Trilobite fragments in limestones. Forms base of Walcott's 1- to 8-inch bedded oölite zone	2.5	0.8
Interval covered: underlain by gray fissile shale and shaly-bedded limestone. Weathers buff-gray	27	8.3
Limestone: dull-gray, fine-grained, massive but thin-bedded, thicker-bedded up section; slightly argillaceous and finely arenaceous. Weathers drab-gray and forms upper part of "pillowy" cliffs	13	4.0
Sandstone and limestone: dull-green-gray, fine- to medium-grained, thin- and lumpy-bedded, massive sandstones and some thicker, tan weathering beds (1 to 1¾ inches); worm trails on some beds. Sandstone becomes calcareous and passes imperceptibly into blue-gray, nodular, argillaceous and finely arenaceous limestone in upper part which weathers duller- but greenish-gray as lower standstones, and all form rounded (pillowy) cliffs maturely dissected along joint planes; undercut in upper part	45	13.8
Shale and sandstone: dull- and greenish-gray, finely micaceous, fissile shale, and some intercalated fine-grained shaly sandstones		

	Feet	Meters
in lower part. Shale becomes arenaceous upward, passing into dull-green, irregular and shaly-bedded sandstone which becomes more massive-bedded and weathers dull-gray	14	4.3
Limestone: tan overlain by gray, green, and maroon, fine-grained, banded limestone, in hemispheric areas which simulate algal masses. Small lenses of dark-green glauconite in upper beds. Intraformational conglomerate locally developed. Weathers brown and gray and forms ledge in shale. Contains well-preserved brachiopods and rare trilobite cranidia and pygidia. Fossil loc. C3b	0.5–2	0.15–0.6
Fossils: *Inglefieldia?* sp. *Olenellus?* sp. indet. 2 gen. undet. *Acrothele clitus?* Walcott		
Shale: dull-green, slightly micaceous, very slightly arenaceous, fissile; weathers green and contains a few thin (¼ to 1 inch) beds of limestone which carry rusty and rotten weathered fossils	6.5	2.0
Limestone: dull-blue-gray, coarse- to medium-grained, in 1- to 5-inch beds. Basal 4 feet strongly cross-bedded and contains irregular amounts of sand. Weathers rusty-tan and gray. All beds contain fragments of *Olenellus* but more concentrated and better preserved in upper 2 feet which is fossil loc. C3a	8	2.5
Fossils: *Antagmus* n. sp. *Bonnia* cf. *columbensis* Resser *B. fieldensis* (Walcott) *B.* n. sp. *Olenellus gilberti?* Meek *O.* n. sp. *Onchocephalus thia?* (Walcott) *Micromitra? charon* Walcott		
Limestone and shale: dull-gray, coarsely crystalline, flaggy limestone in 1- to 3-inch beds, and pale-gray, soft fissile shale partings ½ to 14 inches thick. Limestone contains *Olenellus* fragments, and some beds are lenticular. Thinner beds carry worm tubes	5.5	1.8
Limestone and sandstone: dull-tan-gray, coarse-grained, arenaceous limestone; and calcareous sandstone in extremely irregular (lenticular) 2- to 26-inch beds, contains tiny white quartz pebbles or grains. Weathers dull-tan and forms dip slopes in creek bed	8	2.5
Shale and limestone: pale-gray, fissile shale interbedded with dull-tan-gray, coarsely crystalline, arenaceous limestone in 1- to 4-inch beds. Fragments of *Olenellus* in some limestone beds. Interval covered nearly everywhere	5	1.5
Total thickness of Mount Whyte formation	275	84.7

St. Piran sandstone

Sandstone: tan-gray, medium-coarse-grained, thick-bedded, finely banded. Weathers tan and to irregular surfaces. This and 2 underlying intervals form transition zone into overlying basal arenaceous limestone of Mount Whyte formation	2–3	0.6–0.9
Sandstone: tan-gray, medium- and coarse-grained, thick-bedded, calcareous, banded ½ to 1¼ inches with finer-grained, less calcareous sandstone which weathers to darker, raised, lenticular bands. Sandstone weathers brown. Bands simulate chert from distance	8	2.5
Sandstone: tan, brown, and gray, calcareous, medium-grained, in 4- to 14-inch beds. Weathers dull-tan and slightly flaggy *Note:* Here the St. Piran is faulted against the upper part of Mount Whyte limestone by low angle, normal fault which dips west. Section traced and tied in on	6	1.8

	Feet	Meters
upper banded sandstone bed to "Boulder Pass" west of Ptarmigan Lake, and continued northward up south side of Ptarmigan Peak. The St. Piran cannot be accurately tied in at "Boulder Pass" and is subject to error of 5 to 50 feet of thickness.		
Sandstone and shale: tan-gray flecked blue-gray on fresh fracture, medium-coarse-grained, quartzitic, thick-bedded, slightly calcareous sandstone, strongly banded pale-tan. A few gray shale intervals as below	40	12.3
Quartzite and shale: pale- and buff-gray, coarse-grained, arenaceous quartzite in beds ⅓ to 3½ feet thick, interbedded with dull-gray, argillaceous fissile shale in zones 2 inches to 2 feet thick. Excessive, rusty-brown grains of limonite in upper fourth. Weathers buff-gray and forms slope below Ptarmigan and Mount Whyte cliffs on east spur of Ptarmigan Peak. Slickensides on some bedding surfaces in upper part of interval	118	36.4
Quartzite, shale, and sandstone: as in underlying interval, but much grayer fissile shale; and fewer but thicker beds of quartzite	63	19.4
Quartzite, shale, and sandstone: as in underlying interval, but much more shale, and quartzite more strongly cross-bedded. Worm tubes (*Cruziana*?) and trilobite tracks common on some of thin-bedded sandstones	49	15.1
Quartzite, shale, and sandstone: quartzite as in underlying interval, and pale- and tan-gray, fissile shale partings, and some intercalated platy sandstones. Quartzite strongly cross-bedded and banded	56	17.2
Quartzite, sandstone, and shale: quartzite as underlying, interbedded with brown-gray, thin-bedded sandstone and some dull-tan-gray, fissile, slightly arenaceous shale. Tiny white quartz pebbles; some banding; and cross-bedding in thick quartzite beds. Interval weathers tan-brown and forms cliffs	54	16.6
Quartzite: red and brown banded, coarse-grained, cross-bedded. Weathers dark-tan-brown stained red and forms large angular blocks on bench above lower cliffs	52	16.0
Quartzite: similar to underlying, but more buff on fresh fracture, and argillaceous and finely arenaceous. Weathers gray with pinkish and buff casts	51	15.7
Quartzite: pale-gray and some buff-gray on fresh fractures. Contains 1/16- to ¼-inch pebbles of white quartz in some lower beds. Weathers tan-gray stained deep-brown over large areas	143	44.0
Quartzite: dull- and pale-gray, fine- to medium-grained, hard, massive, in ½- to 2½-foot beds. Weathers tan or gray stained tan and forms ledges and cliffs. Occasional bed pale-red	72	22.0
Sandstone: similar to underlying. Nearly all thin-bedded, banded, quartzitic sandstone. Forms upper part of slope below tan-brown weathered cliffs	41	12.6
Quartzite and sandstone: alternate zones, gray, thicker-bedded, hard quartzite, and thicker zones of tan-gray weathering, thinner-bedded sandstone. Occasional pebbles of white quartzite in upper, coarse-grained beds, and some purple banding and cross-bedding in coarser-grained, thicker quartzite beds. Interval forms slope above Lake Louise shale on west side of saddle	59	18.1
Total thickness of St. Piran sandstone	815	250.6

Fort Mountain sandstone (type locality)

Lake Louise shale member

Shale and sandstone: maroon, pale-green, drab-tan, and brown, arenaceous and some pure shale in alternate zones, all interbedded with drab-tan, brown, and gray, fine- to medium-

	Feet	Meters
grained, thin-bedded sandstone which contains worm tubes and trilobite trails on some beds in upper part. Green shale in upper 6 feet only. In middle third, several zones of ¼- to ¾-inch beds of smooth brown weathering sandstone, and in lower third, dirty-gray, lenticular and nodular-bedded, rusty red-brown-weathering sandstone. Green and tan shale partings between some sandstone beds	60	18.5
Total thickness of Lake Louise shale member	60	18.5
Note: In measuring, the section was traced south and down the dip-slope formed by the upper bed of the Fort Mountain sandstone to the saddle. The above part of the section was measured westward across the saddle and up the east spur of Ptarmigan Peak proper.		
Quartzite and sandstone: tan-gray, coarse- to medium-grained, arenaceous quartzite in 2- to 10-inch beds; contains large worm tubes and *Scolithus*. Weathers buff-tan. Forms top of hill east of east spur of Ptarmigan Peak. Some green and tan shaly partings in upper few feet	16	4.9
Quartzite: white-gray, in beds 1 to 5 feet thick; contains pale-tan limonite grains; weathers white-gray stained buff and forms cliffs. Occasional intercalated, white-gray, coarse-grained sandstone in 1- to 4-inch beds. Much covered with black lichen	82	25.2
Quartzite: white-gray, slightly vitreous, thick-bedded. In lower third, more arenaceous and contains *Scolithus*. Weathers white-gray and forms lower part of white-gray cliffs	84	25.8
Sandstone, quartzite, and shale: tan-gray, quartzitic sandstone, in beds ½ to 3 feet thick; interbedded with drab sandstone and arenaceous shale ¾ to 1 inch thick. *Cruziana* on sandstone and *Scolithus* borings in 3-foot quartzitic sandstone in upper part of interval. Weathers dark-brown and forms transition zone or top part of dirty interval below white-gray quartzitic cliffs	70	21.5
Similar to underlying but contains more and thicker-bedded quartzite. Forms part of dirty-brown zone	108	31.2
Sandstone and shale: dark-tan-gray and gray, coarse-grained, quartzitic sandstone in ½- to 14-inch beds; dark-brown sandstone; and 2-inch to 3-foot zones of black-brown, slightly fissile, finely arenaceous, interbedded shale. Interval weathers black-dirty-brown and forms slopes except in cliff faces of young cirques	60	18.5
Quartzite and sandstone: dull-pale-gray and brown, thick-bedded quartzite and interbedded gray and tan banded, medium-grained sandstone. Weathers dirty-brown	58	17.8
Quartzite: same as underlying but slightly thinner-bedded and forms rounded slope above cliffs	35	10.8
Quartzite: white-tan- and dull-gray, fine- to medium-grained, hard, slightly vitreous, in 4-foot beds; generally smooth-bedded; weathers rusty-tan and forms lower rusty cliffs of Fort Mountain quartzite	118	36.3
Shale: black-gray and dull-tan-gray, hard, chunky to fissile, and some interbedded thin sandstones. Lower part weathers black-gray; upper part bright-tan below rusty-brown cliffs	42	12.9
Sandstone: drab-green, gray, and tan, medium-grained, generally thin-bedded. Contains some tiny quartz pebbles in lower part. Weathers drab-green-gray stained tan. Several zones of shaly or thin-bedded, intercalated sandstone	17	5.2
Conglomerate and sandstone: tan-gray conglomerate in beds ⅔ to 2 feet thick. Pebbles of milky quartz ⅛ to 2¼ inches in diameter, average ⅝ inch in lower part and ¼ inch in upper. Pebbles stained rusty-tan with limonite. Beds weather gray. Interbedded drab-green-tan, thinner-bedded, finer-grained, mica-		

FIGURE 1. ALGAE AND CALCITE VEINLETS IN LOWER PART OF PTARMIGAN LIMESTONE, CASTLE MOUNTAIN SECTION

FIGURE 2. VESICULAR WEATHERING IN LOWER PART OF ELDON DOLOMITE, CASTLE MOUNTAIN SECTION

FIGURE 1. WEST FACE OF PTARMIGAN PEAK SHOWING UPPER PART OF SECTION
P=Pika formation, E=Eldon dolomite, S=Stephen formation, C=Cathedral dolomite, Pt=Ptarmigan limestone, M=Mount Whyte formation.

FIGURE 2. EAST FACE OF PTARMIGAN PEAK SHOWING PIKA FORMATION ON ELDON DOLOMITE
P=Pika, E=Eldon, S=Stephen, C=Cathedral, Pt=Ptarmigan, M=Mount Whyte, SP=St. Piran formations

PTARMIGAN PEAK

	Feet	Meters
ceous sandstone which contains larger grains of subangular white quartz; weathers drab-green and forms 4- to 24-inch zones between conglomerate beds. Upper 8 to 10 feet tan-gray, coarse-grained, quartzitic sandstone which contains white quartz pebbles 1/16 to 5/8 inch in diameter on bedding surfaces, and bright-tan limonite grains	59	18.1
Shale: black-green-gray, hard, slightly metamorphosed, thickly fissile, finely arenaceous. Several 1/4- to 1-inch beds in 6- to 8-inch intervals of finely banded rusty-weathering sandstone. Shale weathers black-gray; some beds stained rusty-tan, and all smoother than the Hector shale	32	9.8
Sandstone and conglomerate: drab-green-tan and gray, coarse-grained, quartzitic sandstone in 4- to 18-inch beds, interbedded with tan and gray sandstone conglomerate which contains pebbles of milky, glassy quartz 1/16 to 1 1/4 inches and average 1/4 inch in diameter in lower beds and 5/8 inch in upper beds. Sandstones contain much limonite and tiny quartz pebbles. All pebbles irregularly arranged (not in lenses) and well rounded. Beds weather drab-rusty-gray stained brown and form ledges	24	7.4
Total thickness of Fort Mountain sandstone (Lake Louise shale not included)	805	245.4
Total thickness of Lower Cambrian	1955	599.2
Total thickness of Ptarmigan Peak section	4730	1452.8

Note: The above part of the section was measured on the ridge east of Ptarmigan Peak from a point approximately 8300 feet in elevation, almost 1/4 mile west of "Deception Pass," westward over 2 low hills and on up the cliffs to the fault on the east spur of Ptarmigan Peak proper.

COMPARISON OF ORIGINAL AND EMENDED PTARMIGAN PEAK SECTIONS

A comparison of the original and emended sections measured on Ptarmigan Peak (Fig. 3) shows six significant errors in the original section [4] of Walcott (1928, p. 278-281).

Walcott assigned all the rocks above the Ptarmigan limestone to the Cathedral formation and estimated their thickness as 2100 feet. The 620 feet of rocks immediately above the Ptarmigan limestone is typical, massive, thick-bedded, buff and gray Cathedral dolomite. Above the Cathedral is 315 feet of Stephen shale and limestone closely similar to that on Castle Mountain. The basal 30 feet of these beds contains a prolific *Glossopleura* fauna, characteristic of the lower Stephen throughout this part of the Canadian Rockies. These 315 feet of limestones and shales on Ptarmigan and Pika peaks (Pl. 6, fig. 1) and on Redoubt Mountain are eroded and weathered into the same form of bench between the underlying Cathedral and overlying Eldon sheer cliffs as on Castle Mountain (Pl. 2) and Mount Bosworth. Above the Stephen on Ptarmigan Peak is 1230 feet of massive Eldon dolomite which forms the sheer cliffs

[4] Walcott's first published section on Ptarmigan Peak was incomplete (Walcott, 1917a, p. 2-5). The one he published in 1928 (Walcott, 1928, p. 278-281) is the original complete section of these rocks and is referred to in this paper as the original section.

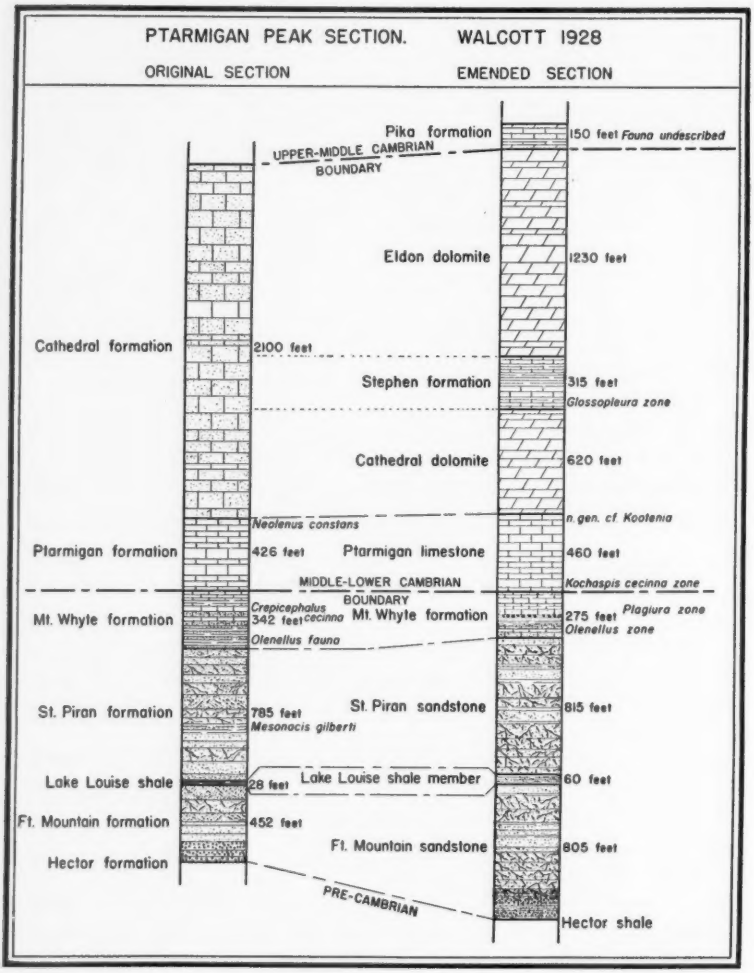


FIGURE 3.—*Comparison of original and emended Ptarmigan Peak sections*

near the top of the mountain (Pl. 6, fig. 2). The Pika formation is also represented by 150 feet of shale and limestone on the northwest crest of Ptarmigan Peak. These beds contain many trilobite cranidia and pygidia and are drag-folded against the massive Eldon dolomite on the north crest of the mountain (Pl. 8, fig. 2). Thus, the upper 2100 feet

of the original Ptarmigan Peak section assigned by Walcott to the Cathedral formation consists of four formations which have a total thickness of 2315 feet.

In the original section the rocks were indicated to be Lower and early Middle Cambrian. The discovery of the Pika and its characteristic fauna proves that Ptarmigan Peak is capped with a thin veneer of early Upper Cambrian sediments and that much more of the Cambrian is represented than was formerly known.

The Ptarmigan formation as originally defined (Walcott, 1917a, p. 1-4) is essentially correct in that the lower boundary was drawn at the base of the black-gray thin-bedded limestone which contains *Kochaspis*, and the upper boundary was drawn at the contact of the black-gray limestone against the overlying buff-weathering Cathedral dolomite. Thus delimited, the Ptarmigan formation on Ptarmigan Peak (the type section) is a natural lithologic unit which is readily recognizable in the field and therefore can be mapped on a scale as small as 3 miles to the inch. The formation is defined in the same way, and the boundaries are drawn at the same horizons in the emended section. Unfortunately, in Walcott's 1928 section (original of figure 3) the Ptarmigan-Mount Whyte boundary was raised 90 feet, thus placing the basal Middle Cambrian beds in the Mount Whyte formation and in the Lower Cambrian. This mistake has caused confusion and errors because such trilobites as *Kochaspis* were assigned to the Mount Whyte and therefore believed to be Lower Cambrian.

The Mount Whyte formation is 67 feet thinner than is shown in the original section (Fig. 3). However, if the 90 feet of beds taken from the Ptarmigan is replaced and its thickness deducted, the thickness obtained by Walcott is approximately that obtained by careful measurement made in 1938.

The Lake Louise shale is more nearly 60 feet than 28 feet thick as given in the original Ptarmigan Peak section—an error of nearly 57 percent. Although the Lake Louise shale is well exposed on the east side of Ptarmigan Peak and Redoubt Mountain and is easily identified in the field, the shale is much too thin to be mapped on a scale as large as one mile to the inch and has never been mapped. Consequently, the Lake Louise shale cannot be designated a formation and is included as the upper member of the Fort Mountain formation in the emended Ptarmigan Peak section (Fig. 3).

The Fort Mountain sandstone in the Ptarmigan Peak section is 805 feet instead of 452 feet thick—an error of 36 percent. As the Fort Mountain sandstone was reported to be at least 940 feet thick on Fairview Mountain (Walcott, 1917a, p. 6) and is 985 feet thick on Castle Moun-

tain (Fig. 2) the abnormal thinness of these rocks on Ptarmigan Peak was inexplicable. That at least 805 feet of Fort Mountain sandstone is present on Ptarmigan Peak is more nearly normal and corrects this puzzling inaccuracy in the literature.

ROSS LAKE AREA, BRITISH COLUMBIA

GENERAL STATEMENT

The original section measured on the cirque wall near Ross Lake contains only a small part of the Cambrian series in this area. Nevertheless, this section is of considerable importance in that it furnishes evidence concerning such significant problems as the stratigraphic position of the *Albertella* zone, the character of the lower Middle Cambrian sediments in the area between Mount Bosworth and Lake Louise, and especially the distribution of the Ptarmigan formation. The Ross Lake section also was the apparent key in the solution of the *Albertella* fauna concerning which many queer and ambiguous statements (p. 1012–1013) and undemonstrable conclusions had been written. In the hope of obtaining accurate data on these problems and laying the groundwork for the later remeasurement of the Mount Bosworth section, the Ross Lake section was selected as the third to be remeasured in 1938.

LOCATION OF AREA

Ross Lake lies approximately seven-tenths of a mile S. 40° W. of Kicking Horse Pass and six-tenths of a mile west of the Continental Divide (Fig. 1). The lake lies at the bottom of a sheer-walled cirque on the north side of an isolated unnamed peak 1 6/10 miles northeast of Narao Peak and 2¼ miles north of Popes Peak. The area is in Yoho National Park near the east edge of the province of British Columbia. The section traverse is 2¼ miles S. 38° E. of Mount Bosworth, 8 miles east of the Burgess shale quarry on Mount Field, and 10 miles S. 70° W. of Ptarmigan Peak.

ROSS LAKE SECTION

The section was measured on the spur southeast of Ross Lake, starting at a point S. 66° E. of Ross Lake at the lowest outcrop of St. Piran sandstone, at an elevation of approximately 6250 feet in the SE. ¼ sec. 34 and the SW. ¼ sec. 35; T. 28 N., R. 17 W. Lower strata crop out on the slope to the east but are separated by covered intervals to the extent that a continuous section is not available below the lowest beds measured.

The upper cliffs on the northwest spur of Narao Peak and along the Continental Divide northeast of Narao Peak and northwest of Mount Niblock are formed of Eldon dolomite and are probably overlain by

Pika shale and dolomite. The bench at the top of the cliffs southeast and southwest of Ross Lake is formed of Stephen shale and limestone. Many pygidia of *Glossopleura* cf. *boccar* (Walcott) from the basal Stephen are present in the talus upon the Cathedral dolomite on the spur southeast of Ross Lake. The Stephen and Eldon were not measured, but the thicknesses appear to be normal and are tentatively estimated to be 1100 feet and 350 feet respectively.

MIDDLE CAMBRIAN

	Feet	Meters
Cathedral dolomite		
Limestone: dark-gray, fine- to medium-grained, thin-bedded, massive, thicker-bedded above. Weathers dark-gray and forms sheer cliffs. Grades laterally to tan-weathering dolomite on the southeast side of the ridge	50–60	15.4–18.5
Ross Lake shale member (type locality)		
Shale and limestone: dark-gray shale as in 1.75- to 2-foot zone below, but upper 16 to 18 inches contains irregular nodules and bands of dark-gray limestone which weathers to cavities stained buff. Shale contains numerous fossils. Top of fossil loc. C5c	6	1.8
Limestone: one bed; dull-gray, nodular, medium-grained. Contains many flakes and blebs of dull-tan clay, and numerous cranidia of *Ptarmingia* and other genera. Weathers buff-gray and to irregularly banded surfaces and forms separation between lower and upper zones of pure Ross Lake shale	0.3–0.5	0.09–0.15
Shale: dark- (black) gray, fine-grained, dense, hard, platy, calcareous, weathers olive-green-gray and breaks down to platy, smooth slabs. Here base of fossil loc. C5c. Forms smooth green-gray cliff or ledge in otherwise tan or gray, rough cliffs of Cathedral dolomite here. Elsewhere usually covered with talus and forms bench	1.75–2	0.5–0.6

FOSSILS: *Albertella bosworthi* Walcott
A. helena Walcott
A. robsonensis Resser
A. similaris Resser
A. 2 n. sp.
"Amecephalus" stator (Walcott)
"A." 2 n. sp.
Kochina bosworthensis Resser
Poliella chilo? (Walcott)
Ptarmingia sp.
Vanuxemella nortia Walcott
Hyolithes cercops Walcott
Acrothele colleni Walcott
Obolus parvus Walcott
Wimanella simplex Walcott

Total thickness of Ross Lake shale member	8.5	2.55
Limestone: dark-gray (lead colored), fine-grained, massive- but thin-bedded (½ to 4 inches); contains flakes and nodules of tan clay. Weathers dark-gray and is usually covered by talus	15–20	4.6–6.2
Dolomite: as underlying, but slightly thinner-bedded. Weathers buff-tan	43.5	13.3
Dolomite: mottled black- and pale-gray as underlying	60	18.5
Dolomite: dark- and light-gray mottled, medium-grained, extremely thick-bedded (2 to 8 feet). Contains coarsely crystalline, white		

	Feet	Meters
dolomite stringers in lower part. Weathers dull-tan and forms cliffs. Occasional 1- to 2-foot zone of dark-gray, finely wavy banded, platy- (¼ to ½ inch) bedded dolomite which weathers dark-gray	57	17.5
Dolomite: pale-gray, fine-grained (nearly lithographic), finely wavy banded, weathers white-gray, and fine bands weather in relief. May represent algal banding	11	3.4
Dolomite: dark- and some white-gray, medium- and coarse-grained, thick-bedded. Weathers dull-tan. Near middle, dark- and light-gray mottled beds	35	10.8
Dolomite: dark-gray, fine- and medium-grained, very slightly petroliferous, massive, thin-bedded in lower, and thick-bedded or unbedded in upper part. Some pale-gray, thick-bedded dolomite in middle part, and some beds slightly banded with white dolomite stringers and veinlets. All weather drab or dark tan-gray above underlying bright-tan beds	57	17.5
Dolomite: pale-gray, medium crystalline, hard, massive, in 1- to 3-foot beds. Weathers buff-tan as underlying beds	72	22.2
Dolomite: pale-gray, fine- to medium-grained, spotted tan. Lower 5 to 8 feet darker-gray, thinner-bedded dolomite. Upper beds massive, 1 to 4 feet thick. All weather bright-tan and form cliffs. Closely similar to basal part of Cathedral dolomite on Ptarmigan Peak	56	17.2
Total thickness of measured Cathedral dolomite (Walcott's Ptarmigan)	480	147.65
Total thickness measured of Middle Cambrian	480	147.65

LOWER CAMBRIAN

Mount Whyte formation

Limestone: gray, blue-gray, and tan, thin-bedded, massive, finely to medium crystalline, and occasional 4- to 8-inch bed of oölite. Upper beds wavy banded, darker-gray, very slightly vitreous. Weathers to bright-tan-buff, platy beds, and forms cliffs. Contains many partings and flakes of tan clay and occasional cranidia of *Alokistocare?* sp., *Zacanthoides? sp.*, and several other genera similar to those in fossil loc. C3c. This is fossil loc. C5b	26	8.0
Limestone: dull- and blue-gray, finely crystalline, massive but platy-bedded in upper part, thick- (4 to 8 inches) bedded in lower part. Weathers buff-tan and to irregular plates and forms cliffs	56	16.9
Limestone: dull- and blue-gray, finely oölitic, hard, thick- and thin-bedded (6 to 1 inches), extremely massive. Oölites slightly coarser in upper beds. Some beds dark-tan-gray	22	6.8
Note: Above interval seems to be same as the "oölite zone" in Ptarmigan Peak section but contains no shale, whereas shale forms approximately 40 percent of the interval on Ptarmigan Peak.		
Shale, sandstone, and limestone: buff and greenish-gray shale replaced upward by arenaceous shale and some shaly sandstone in turn replaced by dull-gray, argillaceous and finely arenaceous sandstone. Limestone increases in amount upward and becomes purer. *Cruziana* tubes present on some beds. All weather tan-gray and form cliffs. Beds massive and much metamorphosed but similar in composition to those beneath oölite interval on Ptarmigan Peak	62	19.1

	Feet	Meters
Shale and limestone: shale in lower part as in underlying interval; becomes more buff-gray upward and interbedded with gray calcareous shale and partings of thin argillaceous limestone; weathers buff and gray banded and forms cliffs	45	13.8
Shale and limestone: dull-gray, and upward, green-gray, fine-grained, hard, argillitic shale, which is calcareous in some beds. Weathers drab-green-gray and to fissile plates. Largely covered by soil and talus	53	16.3
Limestone: dull- and tan-gray, medium-grained, thin- and irregular-(flaggy) bedded (1 to 4 inches). Contains extreme amounts of tan limonite, and poor cranidia of *Bonnia* sp., and *Olenellus* sp. concentrated in lenticular nodules. Fossil loc. C5a. Greenish, thin shale partings between some beds. Weathers tan-gray and forms cliffs or ledges	21	6.5
FOSSILS: *Bonnia* cf. *fieldensis* (Walcott) *Olenellus* sp. *Syspacephalus* n. sp.		
Shale, limestone, and sandstone: green-gray, argillitic, hard shale, interbedded tan- and some maroon-gray, arenaceous limestone, and a few beds of tan calcareous sandstone. Contains a few fragments of *Olenellus* sp. near base, and worm tubes on some shaly sandstones. Interval covered nearly everywhere but well-exposed in gully bottom here	30	9.2
Total thickness of Mount Whyte formation	315	96.6
St. Piran sandstone		
Sandstone: maroon-tan and purple banded, medium- and coarse-grained, in ½- to 2-foot beds; weathers maroon and to thin beds. Upper 5 feet one bed of tan-gray calcareous massive sandstone; weathers tan	51	16.2
Interval covered: probably underlain by thin-bedded, more or less calcareous sandstone, and green-gray argillitic shale	38	11.7
Sandstone: dull-dark-maroon and some red-tan, medium-grained, dense, very slightly calcareous, in 8- to 18-inch beds. A few green, slickensided shale partings in upper part. Top 2½-foot bed weathers tan. Lower third of interval covered	40	12.3
Sandstone: tan, maroon, and some gray, medium- and fine-grained, in 1- to 2-foot beds. Slightly calcareous, weathers gray and dull-maroon and to rounded surfaces	15	4.6
Quartzite and argillite: maroon-gray and tan, coarse- to medium-grained, slightly arenaceous and calcareous quartzite in 10- to 24-inch beds, interbedded with dull-green, finely arenaceous argillite, in 1- to 2-foot zones. Near base one 10-inch zone of red hematitic sandstone. Beds weather dull- and green-gray and are folded into small monocline here	22	6.8
Argillite: dull-green and green-gray stained tan. Compact, hard shale or argillite banded tan on fresh fracture. Weathers dull-green stained tan	10	3.1
Quartzite: green, maroon, and some gray, coarse-grained, more or less arenaceous, in 10- to 18-inch beds, slightly vitreous. Some maroon beds banded irregularly dark-green. Weathers dull-maroon. Upper 1 foot tan-gray, thin- and flaggy-bedded; weathers gray	19	5.8
Total measured thickness of St. Piran sandstone	195	60.5
Total measured thickness of Lower Cambrian	510	157.1
Total measured thickness of Ross Lake section	990	304.75

COMPARISON OF ORIGINAL AND EMENDED ROSS LAKE SECTIONS

In the original Ross Lake section (Walcott, 1917b, p. 13-16) the 640 feet of beds above the Mount Whyte formation was referred to the Ptar-

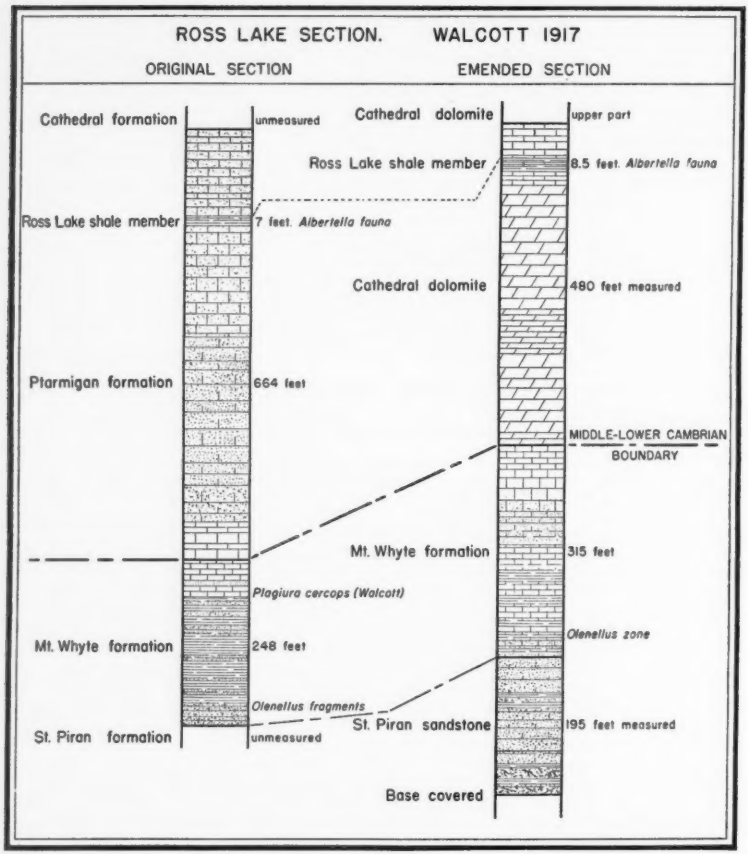


FIGURE 4.—*Comparison of original and emended Ross Lake sections*

migan formation and described as arenaceous limestone,[5] calcareous sandstone, and limestone. In reality, all the rocks below the Ross Lake

[5] Walcott nearly always used the terms "arenaceous limestone" or "calcareous sandstone" in referring to dolomites, particularly to thin-bedded or platy ones. As years of field experience in the Cordilleran region were needed to discover this generalization, it is noted here in order that future workers may be aware of this habitual error in many of Walcott's publications on the Cambrian rocks of this part of North America.

shale, except the underlying 20 feet, are tan and buff-weathering, hard, thick- and thin-bedded, unfossiliferous dolomites characteristic of the Cathedral formation in all the surrounding sections. A change in color or lithology also does not occur at the horizon at which Walcott drew the upper boundary of his Ptarmigan formation in the Ross Lake section. Consequently, the Ptarmigan formation is not recognizable at Ross Lake, and, as the upper and lower dolomites of the Cathedral formation cannot be separated, the 640 feet of beds assigned to the Ptarmigan does not comprise a formation. The *Albertella* fauna is now known to occur in the Cathedral 200 feet above the base of the Ptarmigan on Castle Mountain (Fig. 2) but is not present in the type section of the Ptarmigan limestone on Ptarmigan Peak. Therefore, the occurrence of the *Albertella* fauna in a shale 430 feet above the base of the dolomite at Ross Lake and Mount Bosworth is excellent, if not irrefutable, evidence that the dolomite belongs to the lower part of the Cathedral and that the Ptarmigan formation is not present in this section.

In light of these facts a comparison of the original and emended Ross Lake sections (Fig. 4) shows the Cathedral dolomite resting upon the Mount Whyte, the Ptarmigan formation absent, and the Ross Lake shale a member of the Cathedral dolomite. Walcott (1917b, p. 16) stated that the *Albertella* zone is 500 feet above the Mount Whyte at Ross Lake, whereas the zone is approximately 430 feet above the Mount Whyte in this section. This horizon is more nearly that of the Ross Lake shale on Mount Bosworth 2 miles to the north where the *Albertella* shale, according to Burling (1916b, p. 470), is 375 feet above the top of the Mount Whyte.

The other differences in the original and emended sections are of less significance than those just noted. The Mount Whyte formation is 315 feet instead of 248 feet thick as given in the original section, and the 195 feet of St. Piran sandstone exposed continuously beneath the base of the Mount Whyte formation is shown in the emended section.

CAMBRIAN FORMATIONS

GENERAL STATEMENT

Revision of formations is an exacting and difficult task even under simple field conditions in which the lithologic units are sharply delimited and contain many widely distributed fossils in narrowly restricted zones. Probably, the most important requirements for such work are accuracy of observation in obtaining field data, maturity of judgment in the interpretation of the data, and uniformly close adherence to the established rules which govern sound classification and nomenclature of rock units. In emending the Lower and Middle Cambrian formations in southwestern

Alberta and southeastern British Columbia the writer has tried to follow closely those rules or procedures which have come to be accepted by geologists of the United States Geological Survey and others who have had their major experience in field work (Ashley *et al.*, 1933, p. 423-447).

The prime requirement which a rock unit (sedimentary, igneous, or metamorphic) must satisfy is mappability if it is to be of use to structural or economic geologists, stratigraphers, petrographers, or students of historical geology. The all-significant fact must be clear (but apparently is not always) that unless a rock unit can be graphically portrayed on an areal map such a unit can have little value as a building block in the ultimate geologic history of any given region.

Many of the older formations in Alberta and British Columbia were defined for purely stratigraphic purposes, either as convenient artificial subdivisions of measured sections or as a convenient way to designate strata known to contain a certain assemblage of fossils. The upper and lower boundaries of such "formations" were drawn at different horizons in different sections, usually without regard to lithic similarity or differences. In several extreme cases even the faunas were ignored or never discovered, and the rocks were thus incorrectly assigned to loosely defined "formations" which contained fossils elsewhere.

The usefulness and adaptability for cartographic purposes of the emended formations were adequately checked by mapping them with a plane table and open-sight alidade on scales of 1 and 3 miles to the inch in the Castle Mountain and Ptarmigan Peak areas. Except for the Eldon dolomite and the St. Piran and Fort Mountain sandstones, the formations emended in the following pages all contain readily recognizable faunas and satisfy all the other important requirements set up by the committee of the Geological Society of America on classification and nomenclature of rock units (Ashley *et al.*, 1933, p. 423-447).

FORT MOUNTAIN SANDSTONE

General statement.—The basal Lower Cambrian formation in Banff and Yoho National Parks was originally named the Fairview from Fairview Mountain in the Bow Range, south of Lake Louise (Fig. 1), which was designated the type locality (Walcott, 1908a, p. 5). As the name Fairview was preoccupied, Walcott (1912b, p. 131, footnote a) changed the name to "Fort Mountain sandstone" and the type locality to Redoubt Mountain, formerly known as Fort Mountain (Walcott, 1928, p. 254, footnote 2). Unfortunately, Walcott never published a measured section of these rocks on Redoubt Mountain but instead he (Walcott, 1917a, p. 5-6) gave a generalized section measured on the east side of Ptarmigan Peak, 3 miles north of Redoubt Mountain. In the same paper he pub-

lished the section of the Fort Mountain sandstone on Fairview Mountain. In 1928, in his last discussion of the formation, Walcott (1928, p. 254-255) again named Redoubt Mountain as the type locality but still did not give a detailed section from the type locality. In the same paper he said: "At Redoubt Mountain the formation is finely exposed but it was not measured in detail." He then mentioned the thickness of the sandstone on Fairview Mountain but failed to refer to his section and the thickness of the formation on Ptarmigan Peak published in 1917. Thus a clearly specified type section has never been published for the Fort Mountain sandstone. The formation is well developed and is completely exposed throughout a distance of several miles on the east side of Ptarmigan Peak where its contact with the Hector shale (pre-Cambrian) also is well exposed. The exposures of the Fort Mountain and the Hector likewise are continuous with those on Redoubt Mountain (Pl. 7, fig. 1). The Fort Mountain sandstone appears to be nearly uniform lithologically throughout this area, although the basal conglomerate may be thicker on Redoubt Mountain than on Ptarmigan Peak.

In light of these facts the type section of the emended Fort Mountain sandstone is that measured on Ptarmigan Peak (Fig. 3) where the formation is easily accessible and the section can be conveniently studied and rechecked by future workers.

The original definition of the Fort Mountain sandstone is (Walcott, 1912b, p. 131, footnote a):

"In 1908 . . . these quartzitic sandstones were described as the 'Fairview formation.' As that name, however, is preoccupied in American nomenclature, and as the lower part of the formation has since been found exposed at several places on the east side of the Bow River Valley, it has been decided to apply the name Fort Mountain sandstone to the whole, from the typical exposures on Fort Mountain. The formation is here composed of 2,700 feet of sandstones similar to the 600 feet of sediments to which the name 'Fairview' was applied, and a basal bed of massive conglomerate nearly 200 feet thick. At this locality the Fort Mountain formation is unconformably underlain by pre-Cambrian rocks."

Aside from the fact that the original definition of the Fairview formation is an inadequate description of the Fort Mountain sandstone, the above quotation also contains other errors and ambiguities such as "The formation is here composed of 2,700 feet of sandstones . . ." In his last revision of the formation, Walcott (1928, p. 254) wrote:

"Massive bedded, cliff-forming, purplish, hard, fine-grained quartzitic sandstones, with bands of siliceous and finely arenaceous shale in lower portion. An arenaceous, quartzitic basal conglomerate occurs in some localities."

This description is much more adequate than his earlier ones but is too generalized to give an accurate picture of the formation and creates the erroneous impression that the Fort Mountain consists largely of

purplish sandstone. Consequently the definition of the Fort Mountain sandstone should be revised.

Emended definition.—The oldest Lower Cambrian formation in the vicinity of Ptarmigan Peak consists of 805 feet of coarse- to fine-grained, thick- and thin-bedded, white-gray, tan-gray, and gray quartzitic and impure sandstones interbedded with arenaceous shales and occasional beds of pure quartzite, and locally possesses a basal conglomerate. In the vicinity of Redoubt Mountain and Ptarmigan Peak (Fig. 1) the formation is divisible into five members of which only the upper has been named. The first member (oldest) consists of drab-green-tan and gray, coarse-grained limonitic sandstone in the lower 24 feet, interbedded with tan and gray sandstone conglomerate in which the pebbles are milky quartz 1/16 to 1¼ inches and average a quarter of an inch in diameter. The overlying 32 feet consists of black-green-gray, finely arenaceous argillitic shale, which in turn is overlain by 59 feet of tan-gray conglomerate in beds as much as 2 feet thick interbedded with drab-green-tan thinner-bedded micaceous sandstone. The upper 59 feet of the basal member consists of black-gray shale and drab-green thin-bedded sandstone. The second member, 153 feet thick, is composed of cliff-forming massive white-tan and dull-gray thick-bedded quartzite. The third member, 296 feet thick, consists of dirty-brown, black-gray, and drab-tan, quartzitic, thick- and thin-bedded sandstone interbedded with thin shale intervals and thick beds of quartzite. These rocks weather dirty-brown and form slopes between cliffs. The fourth member consists of 182 feet of white-gray, thick-bedded, cliff-forming interbedded quartzite and sandstone. *"Scolithus"* borings are common in some of the thicker beds in this and in the upper beds of the underlying member. The fifth member (youngest) of the Fort Mountain sandstone is the Lake Louise shale which is 60 feet thick on Ptarmigan Peak and consists of alternating intervals of maroon, pale-green, drab-tan, and brown, arenaceous and some pure shale, all interbedded with drab-tan and gray thin-bedded sandstones.

The Fort Mountain is generally cliff-forming throughout the area and, in the absence of the Lake Louise shale member, is difficult to separate from the overlying St. Piran sandstone. The Fort Mountain rests apparently conformably on the Hector shale in the type of locality on the east spur of Ptarmigan Peak (Pl. 8, fig. 1) but is reported (Walcott, 1910, p. 426-427) to be unconformable upon the Hector if the entire distribution of the formation is taken into account. Fossils are unknown in the Fort Mountain sandstone other than *Scolithus* and *Cruziana*. Although Walcott (1928, p. 254) reported *Iphidella louise* (Walcott), *Hyolithes*, and

"a fragment of a trilobite" from the Lake Louise on Fairview Mountain, fossils could not be found in these rocks in the Ptarmigan Peak area.

Lake Louise shale member.—The Lake Louise shale was originally defined as a formation 105 feet thick in the type locality consisting of "Siliceous shales." (Walcott, 1908a, p. 5). In 1928 Walcott (1928, p. 254, 281, 298, 303) defined the formation as "Compact, gray, siliceous shale." and again gave 105 feet for the thickness of the shale in the type locality near Lake Louise. In the same paper he recognized the shale only in two other sections, Ptarmigan Peak and Mount Assiniboine where it is 28 feet and 70 feet thick respectively. The unit is not thick enough to be mapped on a scale of 1/62,500 and has not sufficient horizontal extent or fossils to give it special importance. Although in the Ptarmigan Peak area the Lake Louise is thicker and more sharply delimited than the lower shales, it seems to be merely another intercalated shale in the Fort Mountain sandstone. Because of these facts the Lake Louise is considered a member of the Fort Mountain sandstone instead of a separate formation. Wherever exposed in the Ptarmigan Peak-Redoubt Mountain area this shale forms an excellent horizon-marker for mapping the St. Piran and Fort Mountain formations.

ST. PIRAN SANDSTONE

The original definition of the St. Piran formation, "Mainly gray, quartzitic sandstones, with a few bands of siliceous shale," (Walcott, 1908a, p. 5) is inadequate. The emended definition (Walcott, 1928, p. 252) is a better lithologic description of these rocks. The formation was emended as follows:

> "This is essentially a sandstone formation, with some greenish, siliceous and arenaceous shales in its upper portion. The sandstones are more or less quartzitic in the middle and lower parts of the section, and vary in color from light gray to dirty gray, brownish, purplish, and pink."

In the same paper Walcott again indicated Mount St. Piran as the type locality of the formation and stated that the formation is 2705 feet thick there. However, in his original description he gave 2640 feet for its thickness in the type locality (Walcott, 1908a, p. 5) and in 1928 gave 2632 feet (Walcott, 1928, p. 303). These discrepancies become much more significant if the reported thicknesses of the formation in surrounding sections are compared. For example, the St. Piran sandstone on Mount Assiniboine is reported to be 525 feet thick, on Castle Mountain 55 feet plus, on Ptarmigan Peak 785, on Mount Odaray 1500 feet, and on Mount Bosworth 503 feet (Walcott, 1928, p. 277, 280, 298, 303, 305, 314). This great haphazard variation in thickness of the St. Piran is not reasonable and probably does not exist. The formation is at least 985 feet instead of 55

feet plus on Castle Mountain (Fig. 2) and 815 feet instead of 785 feet on Ptarmigan Peak (Fig. 3).

As time did not permit remeasurement of the type section in 1938, the St. Piran sandstone is not emended in this paper. Further, Walcott's (1928, p. 252) revised description of the formation applies in general to the rocks in the complete section of the St. Piran on Ptarmigan Peak. However, fossils other than poorly preserved worm tubes *(Cruziana?)* could not be found in these rocks. In areas where the Lake Louise shale is missing, the lower boundary of the St. Piran appears to be poorly delimited, and consequently the St. Piran may have to be mapped with the Fort Mountain sandstone. If this is true, both of these formations will necessarily become members of a new basal Lower Cambrian sandstone formation.

MOUNT WHYTE FORMATION

General statement.—The Mount Whyte formation is the youngest Lower Cambrian formation in the area and is the only one in which identifiable Waucobian fossils were found by the writer. In fact fossils were not found in strata older than the Mount Whyte. Consequently, for purposes of correlation and paleogeography, the Mount Whyte is the most important Lower Cambrian formation in Alberta and British Columbia.

The type locality given in the original definition of the formation was "Mount Whyte, above Lake Agnes, and eastern side of Popes Peak, southwest of Mount St. Piran" (Walcott, 1908a, p. 4). This statement indicates two different areas. At the same time the formation was defined as "Alternating bands of limestone and siliceous and calcareous shale." In 1928 the type locality was stated to be the "North slope of Mount Whyte and southwest of Lake Louise." (Walcott, 1928, p. 251), which also is two different areas. In the same paper the formation was redefined as "Alternating bands of gray and bluish-black limestone, siliceous and calcareous shales, with some interbedded sandstones near the lower part." This definition does not give an accurate or recognizable picture of the formation.

Although the Mount Whyte was first defined in 1908, the section on Mount Whyte was not published until 1928 (Walcott, 1928, p. 302). Strangely, the thickness of the formation in this section is given as 458 feet whereas on page 251 of the same paper it is given as 386 feet. Likewise, the formation is stated to be 252 feet thick on Ptarmigan Peak but in that section it is shown as 342 feet (Fig. 3). The latter discrepancy is explained by the 90 feet of beds which was erroneously transferred in 1928 to the Mount Whyte formation from the original Ptarmigan formation on Ptarmigan Peak (Walcott, 1917a, p. 1-4). Examination of the

Mount Whyte section indicates that the formation there includes Middle Cambrian limestones in the upper 100 feet or more and does not contain as clearly defined or readily recognizable faunal zones as on Ptarmigan Peak. Walcott not only never specified this section as typical of the Mount Whyte formation but nearly always referred to sections in other areas when he discussed the formation or its fauna (Walcott, 1917c, p. 62-67).

Another source of confusion concerning the Mount Whyte formation is that it was known to contain both Middle and Lower Cambrian fossils (Walcott, 1917c, p. 62-64). This fact has been the cause of assigning incorrect stratigraphic positions and geologic ranges to some of the genera and species in the Cordilleran region. The Mount Whyte formation in the Ptarmigan Peak section, often referred to by Walcott, contains the most accurately zoned fossils at present and exhibits clearly discernible boundaries. Because of these facts the Ptarmigan Peak section (Fig. 1) is designated the type of the emended Mount Whyte formation.

Emended definition.—The Mount Whyte is the youngest Lower Cambrian formation in the area studied and rests conformably upon the St. Piran sandstone and beneath the Ptarmigan limestone in the type locality and on Castle Mountain. At Ross Lake the Mount Whyte is overlain by tan-gray, buff-weathering Cathedral dolomite. The formation is divisible into three unnamed lithologic members in the type locality. The basal member consists of 26 feet of flaggy 1- to 14-inch beds of coarsely crystalline gray arenaceous and pure limestone interbedded with lenticular beds of calcareous pebbly sandstone and thin partings of pale-gray fissile shale. Species of *Olenellus* are irregularly distributed throughout but are numerous in the upper 2 feet where they are associated with many cranidia and pygidia of several species of *Bonnia* and a few of *Antagmus, Onchocephalus,* and *Micromitra.* At Ross Lake (Fig. 4) *Syspacephalus* n. sp. also occurs at this horizon, which is named the *Bonnia-Olenellus* zone and is the oldest trilobite zone found in the area.

The middle member of the Mount Whyte is 106 feet thick and consists of dull-green micaceous and arenaceous shale in the lower 21 feet interbedded with a few intraformational conglomerates and thin micaceous sandstones. The shale becomes more arenaceous up section and is replaced gradually by dull-green, shaly-bedded sandstone which in turn becomes calcareous and massive-bedded and passes imperceptibly into blue-gray, nodular, argillaceous and finely arenaceous limestones interbedded with gray fissile shale partings.

The upper member consists of 143 feet of irregularly alternating intervals of limestone and shale. The limestones are strikingly oölitic in

many beds and are dark-blue-gray, pale-gray, or dull-gray in beds which range in thickness from 1 to 24 inches. Occasional beds contain oölites or concretions which range in diameter from 1 to 8 mm., and other beds contain numerous tan siliceous clay flakes and bands which weather buff. Interbedded pale-gray and green-gray shale intervals are numerous and range in thickness from ½ to 48 inches. Intraformational conglomerate in 1- to 4-inch beds are common in the lower and middle part of this member. Trilobite fragments are present but poorly preserved in many of the limestones, and a massive oölitic bed 18 to 22 feet above the base contains many small trilobites belonging to species of *Antagmus?*, *Inglefieldia?*, *Kochiella*, *Plagiura*, *Plagiurella*, and *Syspacephalus?*. This is named the *Plagiura* zone and is the highest stratigraphic horizon from which recognizable fossils were obtained in undoubted Lower Cambrian rocks.

The upper boundary of the Mount Whyte is drawn at the base of the lead-gray massive thin-bedded limestone which forms the lower part of the Ptarmigan limestone on Ptarmigan Peak and Redoubt Mountain. This horizon is also taken arbitrarily as the tentative Middle-Lower Cambrian contact in this section. The fossil now recognized from the emended Mount Whyte formation at the type locality on Ptarmigan Peak are:

n. gen. of *Amecephalina* Poulsen
Antagmus n. sp.
Bonnia cf. *columbensis* Resser
B. fieldensis (Walcott)
B. n. sp.
Inglefieldia? stephenensis (Resser)
I. sp.
Kochiella crito? (Walcott)
Olenellus gilberti? Meek
O. n. sp.
O. sp. indet.
Onchocephalus thia? (Walcott)
Plagiura cercops (Walcott)
P. n. sp.
Plagiurella cleades (Walcott)
P. n. sp.
Syspacephalus n. sp.
Acrothele clitus? Walcott
Micromitra? charon Walcott

PTARMIGAN LIMESTONE

General statement.—The Ptarmigan limestone has been the most confused and misunderstood Middle Cambrian formation in the Canadian Rocky Mountains. The unfortunate and incorrect assignment of the "*Albertella* fauna" to the Ptarmigan has caused probably more erroneous statements and fallacious conclusions during the past 20 years concerning the fossils and correlation of this formation than any other in the Cordilleran trough.

The Ptarmigan formation was originally defined (Walcott, 1917a, p. 1-4) under peculiar circumstances and with considerable doubt about parts of it. Apparently, the formation was set up solely to provide a name for the strata in which the *Albertella* fauna was believed to occur, but Burling (1916b, p. 469-472) had previously shown that the *Albertella*

fauna was several fundred feet above the base and well within the Cathedral formation. *Albertella* and its associated genera are not present in the type locality of the Ptarmigan formation, where fossils are extremely rare in the formation, and the few which are present occur only in the lower and upper parts. What is even more peculiar, the rocks assigned to the Ptarmigan in the Ross Lake section (Walcott, 1917b, p. 14-15) are tan-gray and buff dolomites typical of the Cathedral formation throughout the area. Walcott was aware of these facts, as is shown in his original discussion of the Ptarmigan formation in which he wrote (Walcott, 1917a, p. 3-4):

> "It is possible that the formation is only a broad local lentile that was deposited in a depression of the Lower Cambrian sea bed. My first field impression was that the Ptarmigan limestone was deposited locally in a shallow basin largely as oölites before the coming of the physical change that produced the great Cathedral limestones."

This statement is essentially correct, and no field evidence could be found by the writer to change it. Further, the absence of the Ptarmigan limestone in the Bow Range sections and the fact that the formation is only 105 feet thick at Castle Mountain (Fig. 2), 15 miles southeast of Ptarmigan Peak, also substantiate the correctness of it. Consequently, Walcott's subsequent attempts to assign rocks in the Bow Range section to the Ptarmigan formation were artificial and erroneous.

The original definition of the Ptarmigan formation is (Walcott, 1917a, p. 2):

> "Arenaceous gray limestones with interbedded bands of thinner bedded, dark bluish-black limestones and some interbedded bands of shale."

The type section of the formation was measured on Ptarmigan Peak and is readily recognizable in the field. In the type section the boundaries were drawn at natural lithic changes, and hence the formation there is a mappable unit. In the first complete section of the rocks on Ptarmigan Peak (Walcott, 1928, p. 278-279), the Ptarmigan-Mount Whyte, and therefore the Middle-Lower Cambrian, boundary unfortunately was raised 90 feet, the rocks in parts "1d and 1e" of the original section being assigned to the Mount Whyte. An explanation or the evidence for this change not only was not made, but the original thickness (516 feet) was again given in the definition of the formation instead of 426 feet as in the emended section (Walcott, 1928, p. 250, 278-279).

In light of these facts the Ptarmigan formation is emended, but its type section remains on Ptarmigan Peak. The emended definition given here is based upon the emended type section (Fig. 3) measured on the south side of Ptarmigan Peak in 1938.

Emended definition.—The Ptarmigan limestone on Ptarmigan Peak is bounded at the base by the buff-gray shale and limestone of the upper

Mount Whyte and at the top by the pale buff-gray-weathering coarse-grained thick-bedded Cathedral dolomite. These boundaries are clearly visible for long distances in the sides of Ptarmigan Peak and Redoubt Mountain (Pl. 7, fig. 1) where the formation is readily mappable on a scale of one inch to the mile.

The Ptarmigan is remarkably homogeneous and consists of dark- and lead-gray, generally finely crystalline, thick- and thin-bedded limestone. The lower 71 feet is composed of thin-bedded (nodular and shaly) limestone which contains many tan and gray clay partings and flakes. Coarse oölites or Algae? (Pl. 5, fig. 1) are present in some beds. Rare and fragmentary specimens of *Kochaspis* cf. *cecinna* (Walcott), *Kochaspis* cf. *gogensis* (Walcott), *Lingulella?* sp., *Micromitra* sp., *Nisusia* sp., and *Hyolithes* sp. occur 12 and 40 feet above the base of the formation.

Most of the Ptarmigan consists of lead-gray and sooty-gray, finely to medium crystalline limestone which is slightly petroliferous in some beds, and a few alternating intervals of pale-gray fine-grained limestone. A 9-foot interval of white-gray-weathering, almost lithographic limestone and two intercalated beds of white-gray limestone-breccia occurs 165 feet above the base and form an excellent horizon-marker in the type section. A new genus similar to *Kootenia* is represented by poorly preserved pygidia approximately 25 feet below the top of the formation. This is probably the zone from which Walcott (1917a, p. 2) reported *Neolenus constans* and *Zacanthoides cimon*. The upper 3 feet of the Ptarmigan consists of pale-gray, slightly lithographic, massive, hard limestone which forms a sharp contact with the basal buff dolomite of the overlying Cathedral.

The Ptarmigan limestone forms the lower black-gray cliffs (Pl. 7, fig. 1) on Redoubt Mountain and Ptarmigan Peak. Fossils are extremely rare in the formation. None could be found on Castle Mountain, even though the beds which Walcott (1928, p. 276) said contained a dwarfed fauna were searched carefully. In the type section on Ptarmigan Peak fossils were found only in the lower and upper parts of the Ptarmigan limestone (Fig. 3). The complete tentative list of these genera identified by the writer is given here.

Kochaspis cf. *cecinna* (Walcott)
K. cf. *gogensis* (Walcott)
n. gen. cf. *Kootenia* Walcott
Hyolithes sp.
Lingulella? sp.
Micromitra? sp.
Nisusia sp.

CATHEDRAL DOLOMITE

General statement.—The Cathedral formation was originally defined as "Massive, arenaceous and dolomitic limestone," (Walcott, 1908a, p. 4), and its type locality given as "Cathedral Mountain and Cathedral Crags,

east of Mt. Stephen and southeast of Mt. Bosworth." When the formation was redefined (Walcott, 1928, p. 249) the type locality was changed to the "South face of Mount Bosworth between Hector and Stephen on the Canadian Pacific Railway, also finely exposed in Cathedral Crags east of Mount Stephen." The definition of the formation was also altered to "More or less arenaceous, thick and thin layers of hard, gray limestone forming fine cliffs where not broken down." Although both definitions are inadequate, the original is more accurate in that it mentioned the dolomitic character of the rocks. In discussion the stratigraphic relations of the formation Walcott (1928, p. 249) stated:

> "The lower boundary is clearly marked in the northern cliffs of Popes Peak at Ross Lake on the south side of Kicking Horse Pass opposite Mount Bosworth. The Cathedral terminates below in a massive bedded rough, arenaceous limestone that rests conformably on a thin-bedded arenaceous mottled limestone of the Ptarmigan formation. This same relation also exists in the eastern cliffs of Mount Bosworth overlooking the Bow River Valley."

As already pointed out, the Ptarmigan formation is not recognizable at Ross Lake, and a superficial reconnaissance also failed to reveal it at Mount Bosworth. Examination of Walcott's photographs (1917b, pls. 2, 3) shows clearly that a formational boundary which could be used for mapping is not present at the horizon at which he drew the lower boundary of the Cathedral at Ross Lake because the rocks are identical lithologically several hundred feet above and below that line.

Ptarmigan Peak and Castle Mountain are the only sections known in which the Cathedral was found to be underlain by black-gray limestone of the Ptarmigan formation. However, throughout the region from Mount Stephen to Ptarmigan Peak and southward to Castle Mountain the Cathedral is lithologically uniform and consists of thick- and some thin-bedded, tan-gray and gray, medium- to coarse-grained dolomite. The most striking characteristics of the formation are that it weathers buff, forms sheer cliffs, and contains uniformly large amounts of magnesium. On Ptarmigan Peak, at Ross Lake, and probably on Mount Bosworth, large irregular intervals, 5 to 110 feet thick, of black-gray, *Girvanella*-bearing limestone are present within the middle and upper parts of the formation. These limestones seem to be undolomitized masses of the original formation, most of which have been altered to light tan-gray pure dolomite by the addition of magnesium.

Another serious error concerning the Cathedral formation was the assignment of the *Albertella*-zone (Ross Lake shale) to the Ptarmigan formation instead of leaving it in the Cathedral where it first was placed correctly by Burling (1916b, p. 469-472). The *Albertella* fauna consists of 12 species which occur 200 feet above the base of the Cathedral on Castle Mountain (Fig. 2), and 18 species were found in the Ross Lake

shale 430 feet above the base of the Cathedral in the Ross Lake section (Fig. 4). Consequently the *Albertella* assemblage must be thought of as the Cathedral fauna. Except for the numerous fossils in this shale zone, the massive dolomites which comprise the Cathedral formation are barren of fossils except *Girvanella?* in a few isolated limestone areas.

The Mount Bosworth section which contains the type of the Cathedral dolomite was not remeasured in 1938. The stratigraphic lithologic and faunal data and the corrected errors relating to the formation obtained from the sections which were remeasured and examined are sufficient to warrant the following revision of the Cathedral formation.

Emended definition.—The Cathedral formation in the region between Ptarmigan Peak, Castle Mountain, and Kicking Horse Pass consists of tan-gray, buff-gray, and some dull-gray, finely to medium crystalline, generally pure dolomite in beds half an inch to 5 feet thick. The rocks are uniformly weathered tan and buff-gray and generally form sheer cliffs or steep slopes throughout the mountains. The formation is 670 feet thick on Castle Mountain, 620 feet instead of 2100 feet thick on Ptarmigan Peak (Fig. 3), and probably is less than 1100 feet thick at Ross Lake and on Mount Bosworth instead of 1595 feet as originally reported (Walcott, 1908c, p. 212). Thus, the formation thickens northwestward from a minimum of 620 feet on Castle Mountain to more than 1000 feet in the vicinity of Kicking Horse Pass. The Cathedral dolomite rests upon the lead-gray Ptarmigan limestone in Castle Mountain and Ptarmigan Peak and upon the upper blue-gray and tan slightly oölitic limestone in the upper part of the Mount Whyte at Ross Lake. The upper boundary of the Cathedral is drawn beneath the basal shale of the Stephen on Mount Bosworth, Ptarmigan Peak, and Castle Mountain (Pls. 2, 3, and 6; fig. 1). The formation thus delimited is a natural lithic unit and is readily mappable on as small a scale as 3 miles to the inch. Although fossils are rare or absent in the dolomites and could not be found at Ptarmigan Peak in the entire formation, at least 18 species are present in the Ross Lake shale at Ross Lake. This shale now becomes a member of the Cathedral formation. Of the 19 species given in the following list 12 were found in the shale within the Cathedral dolomite on Castle Mountain. Those which are known from both areas are preceded by an asterisk.

Albertella bosworthi Walcott
**A. helena* Walcott
**A. robsonensis* Resser
**A. similaris* Resser
**A.* 2 n. sp.
**"Amecephalus" stator* (Walcott)
"A." 2 n. sp.
Kochina bosworthensis Resser
**Poliella chilo?* (Walcott)
Ptarmingia sp.
**Vanuxemella nortia* (Walcott)
V. n. sp.
**Hyolithes cercops* Walcott
Acrothele colleni Walcott
**Obolus parvus* Walcott
Paterina wapta (Walcott)
**Wimanella simplex* Walcott

STEPHEN FORMATION

The Stephen formation was remeasured only on Castle Mountain and Ptarmigan Peak in 1938. In this region, east of Bow Valley, the formation is apparently much thinner (285 to 315 feet) than in the type section on Mount Stephen where it is reported to be 902 feet thick (Walcott, 1928, p. 315-317). Part of this difference is accounted for in the western sections where the *Ogygopsis* and Burgess shale members are included in the upper part of the formation.

The Stephen shale in the Castle Mountain section (Pls. 2, 3) is a clearly demarcated shale and thin-bedded limestone which everywhere forms benches or slopes between the massive sheer cliffs of Cathedral and Eldon dolomites. Consequently the Stephen is an excellent mappable unit. The formation is also rich in well-preserved fossils, which on Castle Mountain occur in relatively restricted zones. In the basal 11 feet of shale and limestone are numerous individuals of *Alokistocare?, Chancia, Glossopleura, Kootenia, Zacanthoides,* and others. This fauna is typical of that in the base of the Damnation limestone of Montana (Deiss, 1938, fig. 2) and is known as the *Glossopleura-Kootenia* fauna (Deiss, 1939, p. 39). Rare specimens of *Glossopleura* occur as far as 63 feet above the base of the Stephen. The next faunal zone, tentatively called the *Ehmaniella* zone, is 82 feet below the top of the formation (Pl. 3) in green and gray calcareous shale and limestone which contains *Ehmaniella, Solenopleurella?,* and *Obolus.* The limestones and shales 30 feet above the *Ehmaniella* zone contain numerous individuals of *Solenopleurella* and a few of *Thomsonaspis.* For convenience this is referred to as the *Solenopleurella* zone but contains an aberrant assemblage of genera which are widely separated in northwestern Montana. The upper 6 feet of the Stephen on Castle Mountain contains a few pygidia of *Olenoides* cf. *serratus* (Rominger) and many cranidia of *Solenopleurella* n. sp. and *Thomsonaspis* n. sp. *Thomsonaspis* is unknown elsewhere in the Cordillera except in the Steamboat limestone (Deiss, 1939, p. 45) of northwestern Montana where it is associated with *Kochaspis upis* (Walcott) in the youngest Middle Cambrian fauna yet known. Thus the Stephen on Castle Mountain contains trilobites which are in faunas separated by 900 to 1200 feet of strata in Montana. The problem becomes more confused when the *Ogygopsis* shale with its typical Pentagon limestone fauna is included at the top of the Stephen. This problem is unsolved, but its present status may be summarized briefly.

The Stephen east of the Bow River does not contain the *Ogygopsis* or the Burgess shale faunas or other shales lithologically similar to those on Mount Stephen and Mount Field respectively. The faunas of the *Ogygopsis* and Burgess shales are quite different from each other and

from those in the Stephen east of Bow Valley. The Stephen fauna generally has been thought of as the *Ogygopsis* shale fauna which is quite different from any of the faunules in the Stephen east of Bow Valley. Until much detailed field work is done at least on Mount Bosworth and Mount Stephen, the Stephen formation cannot be adequately emended, and the chronologic and stratigraphic relationships of the *Ogygopsis* shale to the Stephen formation in the eastern sections cannot be determined. However, the Stephen formation in the area between Castle Mountain and Ptarmigan Peak on the east and Ross Lake and Mount Bosworth on the west is a sharply delimited natural lithologic unit and will remain as a useful mappable formation. Although the fossils from the Stephen have not been carefully studied and described, the following genera and species are recognized in the collections from these rocks in the Castle Mountain and Ptarmigan Peak sections.

Alokistocare? sp.
"Amecephalus" 2 n. sp.
Chancia? 2 n. sp.
Ehmania? sp.
Ehmaniella cf. *Walcotti* Resser
Glossopleura boccar (Walcott)
G. cf. *boccar* (Walcott)
G. bosworthensis? Resser
G. fordensis Deiss
G. nitida Resser
G. cf. *walcotti* Poulsen
G. 2 n. sp.
Kootenia sp.
n. gen. cf. *Glossocoryphus* Deiss
n. gen. cf. *Inglefieldia* Poulsen
Solenopleurella 4 n. sp.
n. gen. cf. *Strotocephalus* Resser
Thomsonaspis 4 n. sp.
Olenoides cf. *serratus* (Rominger)
Zacanthoides sp. indt.
Obolus mcconnelli? Walcott
O. sp.
Hyolithes

ELDON DOLOMITE

General statement.—In the original definition of the Eldon (Walcott, 1908a, p. 3) the type locality was placed on Castle Mountain "one to two miles north of Eldon Switch, on the Canadian Pacific Railway," and the formation was described as "massive, arenaceous, dolomitic limestones, with a few bands of purer bluish gray limestone." The boundaries of the Eldon were not indicated, but its thicknesses stated were "In Mt. Bosworth section, 2733 feet; at Castle Mountain, 2,195." Later the same year (Walcott, 1908b, p. 209) the thickness of the Eldon on Mount Bosworth was given as 2728 feet, and *Ogygopsis* and *Bathyuriscus* and other genera of the *Ogygopsis* shale fauna were stated to occur on Mount Stephen at a horizon 700 feet above the base of the Eldon. Fossils were also reported from shaly limestones 812 feet below the top of the formation on Mount Bosworth.

The Eldon was redefined in 1928 (Walcott, 1928, p. 247) as follows:

"Light gray with some dark gray siliceous and arenaceous limestones, mainly in thick layers grouped in massive bands that form great cliffs on Castle Mountain, Mount Bosworth and Mount Stephen. . . . A band of bluish-gray, thin-bedded limestone occurs towards the top, one about midway, and another about 700 feet (213.4 m.) from the base on Mount Stephen."

The thickness of the Eldon on Castle Mountain was now given as 1905 feet instead of 2195 feet as stated in 1908. The type section was again placed on Castle Mountain but "opposite Eldon." In the Mount Bosworth section (Walcott, 1928, p. 208-209) the fauna previously stated to occur 700 feet above the base of the Eldon on Mount Stephen was omitted. An explanation was not given for the correction of this earlier error. If the Mount Stephen section (Walcott, 1928, p. 315) is examined this fauna is found, astonishingly, in the upper 190 feet of the Stephen formation, but *Zacanthoides spinosus* Walcott is omitted. The formation was again stated to be 2728 feet thick on Mount Bosworth. However, in the definition of the formation (Walcott, 1928, p. 247) the Eldon is stated to be 2840 feet thick on Mount Stephen, but in the Stephen section in the same paper its thickness is given as 1680 feet and on Mount Bosworth (Walcott, 1928, p. 315) 1595 feet. Explanations were not offered for any of these contradictions.

Remeasurement of the Castle Mountain and Ptarmigan Peak sections and observation of the Eldon on Mount Bosworth show conclusively that the formation is much thinner than has been reported, that it is a massive, nearly pure dolomite instead of a limestone, and that fossils are absent or extremely rare in the formation. As the sheer cliffs on Castle Mountain opposite Eldon are unclimbable, cannot be studied closely, and, consequently, a detailed section cannot be measured over them, the type section of the Eldon formation is changed to Helena Ridge (Walcott, 1928, p. 274, pl. 40, fig. 1) on the east side of Castle Mountain where the emended section of these rocks was measured (Pl. 4).

Emended definition.—The most striking characteristics of the Eldon formation are the massive thick-bedded dolomites which form sheer gray cliffs of great heights throughout the mountains. The lower boundary of the Eldon is marked by the dull-gray, shaly limestones at the top of the Stephen, and the upper boundary is marked by the thin-bedded pale gray beds at the base of the Pika formation. As these boundaries are often visible for long distances and as the cliff-forming dolomites of the Eldon are conspicuous throughout the mountains, the formation is readily mappable.

The Eldon on Castle Mountain and Ptarmigan Peak consists of massive, thick- and thin-bedded black- and light-gray, medium to coarsely crystalline dolomite in alternating intervals. Not one bed of limestone was found in these sections within the formation. The black-gray dolomites are all more or less petroliferous, and many are vitreous on fresh fracture. Algae? and poorly preserved oölites occur in the lower and upper parts of the formation, and some beds in the lower and many in

the upper part weather to irregular vesicular surfaces (Pl. 5, fig. 2). Except for unrecognizable Algae *(Girvanella?)* fossils were not found in the Eldon.

The formation is 1015 feet thick in the type section (Fig. 2), 1230 feet on Ptarmigan Peak (Fig. 3), and is estimated to be 1200 or 1300 feet thick on Mount Bosworth. Although Walcott (1928, p. 275) stated that the Eldon is 1905 feet thick on Castle Mountain, in reality the formation as he defined it would be 1565 feet thick if the 550 feet of beds which he included in the Eldon were not separated as the Pika formation.

PIKA FORMATION

General statement.—The Eldon formation as last defined by Walcott and shown in his Castle Mountain section (Walcott, 1928, p. 275) consists of two strikingly different types of lithology. The lower 1325 feet ("1d plus 1e") of his section is massive, thin-bedded, cliff-forming, dark- and light-gray crystalline dolomite, and the upper 580 feet ("1a to 1c") is generally thinner-bedded dolomites and limestones, and, in the lower fourth, one shale interval. The lower 1325 feet (actually 1015 feet) forms a natural lithologic unit which is sharply delimited from the upper unit and is the restricted Eldon dolomite as emended in this paper. The 580-foot interval, really 550 feet thick (Fig. 2), is also a natural lithologic and mappable unit, contains an old Upper Cambrian fauna 93 feet above the base, and is defined as a new formation, the Pika (Pl. 4). As the Eldon cliff-forming dolomite seems to be reasonably uniform in thickness in the sections studied, the great discrepancies in the reported thickness of the formation on Mount Bosworth and Mount Stephen may be accounted for by a variation in thickness of the Pika which was formerly included in the Eldon. However, as the maximum thickness of the Pika is known at present only on Castle Mountain this point must remain conjectural until the other sections are remeasured.

The discovery of an Upper Cambrian fauna (undescribed, but older than the *Cedaria* fauna) in the lower fifth of the Pika is of especial importance as the Upper-Middle Cambrian boundary must be somewhere close to and below this zone. The Pika-Eldon contact is believed tentatively to be this boundary.

Original definition.—The type section of the Pika formation was measured on the west side of Helena Ridge (Fig. 1), starting at a point approximately 8300 feet in elevation at the top of the Eldon dolomite cliffs east of the upper cirque-lake on the east side of Castle Mountain (Pl. 4). The name of the formation is taken from Pika Peak which is approximately a mile north of Ptarmigan Peak in the northeast part of the area (Fig. 1).

The lower 93 feet of the Pika formation consists of dark-, blue-, and pale buff-gray, thin-bedded (½ to 5 inches) limestone and pale-gray, thin-bedded, argillaceous dolomite. Poorly developed mud cracks and edgewise conglomerates are present in some of the dolomites. The limestones are generally fine-grained and shaly-bedded and contain oölites and tan arenaceous clay flakes and partings. These limestones are strikingly different in appearance and composition from the underlying Eldon dolomite. The second interval of the Pika, 38 feet thick, is composed of shale, limestone, and intraformational conglomerate. The shale is dull-brown-gray and fissile, and intercalated in it are thin-bedded limestones which contain numerous cranidia and pygidia belonging to a new basal Upper Cambrian fauna (Christina Lochman, personal communication). The upper 20 feet of this member is composed of thin-bedded, wavy banded, blue-gray limestone. The third interval consists of 102 feet of dull- and white-gray, yellow-tan-weathering, thick- and thin-bedded, massive, buff-banded dolomite. The upper interval of the Pika formation, 306 feet thick, consists of dull- and dark-gray, thin-bedded, banded and occasionally mottled "salt and pepper" dolomite which becomes thicker-bedded up section and contains varying amounts of arenaceous buff clay flakes. This interval forms the dull- and light-gray cliffs beneath the maroon slopes of Arctomys shale on Helena Ridge (Pl. 4) and Castle Mountain.

Although the Pika has not yet been measured in the Mount Bosworth area, it is a readily mappable formation in the vicinity of Castle Mountain and Ptarmigan Peak. In the latter section only 150 feet of the formation has not been removed by erosion and forms the north side of the crest of Ptarmigan Peak (Pl. 6, fig. 2).

ARCTOMYS FORMATION

On the north end of Castle Mountain and forming much of the crest of Helena Ridge (Pl. 2) is 270 feet of dark-maroon, buff-tan, dull-gray, arenaceous and dolomitic chunky shales, thin-bedded arenaceous argillaceous dolomites, and pale buff-gray limestones. These rocks weather maroon and bright-tan and form shaly slopes between the underlying Pika and overlying Bosworth cliffs. The lower shales contain pseudomorphs of salt crystals as much as three-fourths of an inch in length and a few mud cracks. Fossils are absent from most of the beds, but poorly preserved *Obolus?* sp. occur 133 feet below the top of the formation. These rocks comprise an unusually natural and sharply demarcated lithologic unit which is generally well exposed and readily mappable on a scale of 1/62,500. These sediments on Castle Mountain were assigned by correlation to the Arctomys formation whose type section is on Sulli-

van Peak in the Glacier Lake area 65 miles northwest of Castle Mountain (Walcott, 1928, p. 245-246). The name Arctomys formation is tentatively accepted for these rocks in the vicinity of Bow Valley, as the formation occupies approximately the same stratigraphic position in its type section.

The Arctomys was first named and defined in 1923 by Walcott (1923, p. 462) as follows:

> "Bluish-gray irregularly laminated cliff-forming limestones which are more or less magnesian in some layers. This limestone is underlain by a series of arenaceous and siliceous shales with bands of hard, finely laminated, dove-colored limestone."

In this paper Walcott (1923, fig. 24) correlated the Arctomys of the type locality on Sullivan Peak with the lower part of the Bosworth formation on Mount Bosworth. Then, in discussing the Arctomys in his last paper, Walcott (1928, p. 245-246) said:

> "Finely laminated, smooth-surfaced limestones overlying siliceous, arenaceous, and calcareous shales and interbedded laminated limestones; the entire formation at Glacier Lake conveys the impression that it was a deposit of fine silts, sands, and calcareous muds or slimes in a shallow sea that marked the beginning of the Upper Cambrian in the Cordilleran Geosyncline in this area.
>
> "The formation appears to represent the period of deposition of a series of shallow fresh-water deposits, alternating with brackish water and marine sediments such as would occur in a shallow sea near the mouth of a large river, bringing fine sand, mud, and slimes derived from low, old land surfaces. These fine shales and sandstones alternate with more or less thin, calcareous and arenaceous layers and have on their surface ripple marks, mud cracks, and casts of salt crystals. They represent, in the Bow Valley area, the period of transition from the massive Middle Cambrian Eldon limestone beneath to the quite different magnesian limestone of the Upper Cambrian Bosworth formation above."

Aside from the facts that the Eldon is dolomite and the Bosworth consists largely of limestone, the Arctomys formation on Castle Mountain overlies the Pika formation, now known to contain a basal Upper Cambrian fauna. Consequently, "in the Bow Valley area" the Arctomys cannot represent the period of transition from the Middle to the Upper Cambrian.

BOSWORTH FORMATION

The 430 feet of rocks forming the highest points on Helena Ridge and the top of the Castle Mountain section is assigned to the Bosworth formation (Walcott, 1928, p. 243-244). As the Mount Bosworth section was not remeasured this formation cannot be emended at present.

On Castle Mountain (Fig. 2) the rocks assigned to the Bosworth consist of a basal 84-foot interval of tan-weathering massive dolomite and alternating zones of green and gray-weathering limestone and shale. The upper 346-foot unit consists of black- and dark-gray, fine-grained, more or less fetid (petroliferous), massive limestones, and a few intervals as much as 20 feet thick of dull-gray laminated dolomite. Some of the lime-

FIGURE 1. NORTH FACE OF REDOUBT MOUNTAIN
S=Stephen, C=Cathedral, Pt=Ptarmigan, M=Mount Whyte formations.

FIGURE 2. CAMBRIAN-BELTIAN CONTACT ON EAST SPUR OF PTARMIGAN PEAK

REDOUBT MOUNTAIN, AND CAMBRIAN-BELTIAN CONTACT

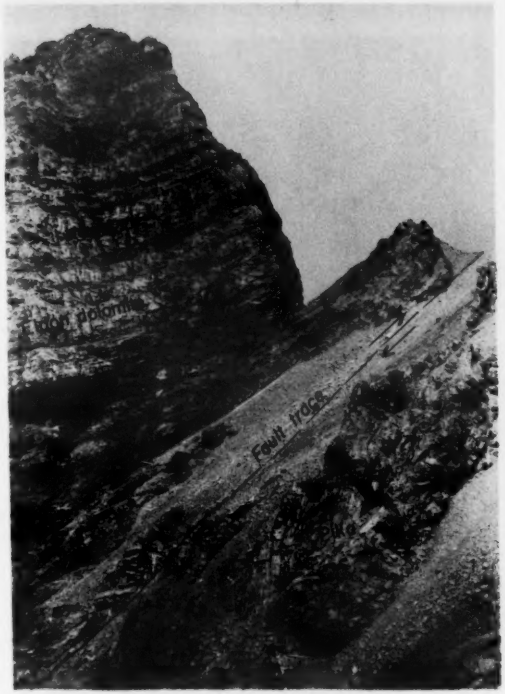


FIGURE 1. PIKA LIMESTONE AND SHALE FAULTED AGAINST ELDON DOLOMITE, NORTH END OF PTARMIGAN PEAK

FIGURE 2. CONGLOMERATE IN HECTOR (PRE-CAMBRIAN) SHALE EAST OF PTARMIGAN PEAK
Length of hammer-handle 15 inches.

PIKA FORMATION, AND CONGLOMERATE IN HECTOR SHALE

stones are finely oölitic and an occasional bed contains *Girvanella*. Some beds of dark blue-gray massive limestone have hemispheric upper surfaces and may be algal reefs (bioherms). Such beds or masses occur 45 feet, 260 feet, and 298 feet above the base of the formation. On Castle Mountain the Bosworth limestone forms sheer or broken dark-gray cliffs above the maroon and bright-tan shaly slopes of the Arctomys.

CAMBRIAN—PRE-CAMBRIAN BOUNDARY

Of the three sections remeasured in 1938 (Figs. 2, 3, 4) the Ptarmigan is the only one in which the basal Lower Cambrian strata are exposed. On the east side of Ptarmigan Peak the weathered contact of the Fort Mountain sandstone and the underlying Hector shale is exposed for long distances (Pl. 7, fig. 2). The basal drab-green-gray, coarse-grained, pebbly sandstone of the Fort Mountain rests in apparent conformity on the underlying Hector shale, which contains beds of intraformational conglomerate whose angular and rounded pebbles are gray, maroon, and green, fine-grained limestone more than 18 inches in diameter (Pl. 8, fig. 1). Approximately 300 feet below the Cambrian-Beltian contact are several beds of intraformational conglomerate which contain flat pebbles of gray fine-grained limestone. On the upper surfaces of the conglomerates are white quartz pebbles ⅛ to ⅜ of an inch in diameter which are similar to those in the basal Fort Mountain sandstone. Likewise, the upper Hector shale is somewhat similar lithologically to the 32-foot interval 24 feet above the base of the Fort Mountain. Although the basal pebbly sandstone ("conglomerate") of the Fort Mountain clearly marks the upper boundary of the Hector shale (Pl. 7, fig. 2), nevertheless the Hector is partly transitional lithologically into the overlying Fort Mountain. Walcott (1910, p. 426) noted that the Fort Mountain and Hector are approximately conformable and pointed out the lithologic differences between the Hector and the shale in the lower part of the Fort Mountain ("Fairview") formation but he failed to note the lithologic similarities exhibited by the two formations. On the other hand he gave convincing evidence which indicates a stratigraphic break between the two formations and which he considered sufficient to establish the pre-Cambrian age of the Hector. In this connection Walcott (1910, p. 426) said:

> "Viewed in a restricted way, much of the pre-Cambrian surface was regular and the Cambrian rocks appear to be conformable to the subjacent pre-Cambrian strata, but when we apply the test of the varying thickness of the basal Cambrian conglomerate and the difference in the character of the upper beds of the Algonkian in different places, we at once become aware that the pre-Cambrian surface is more or less irregular, and that when the Cambrian sea transgressed over the area now included in the Bow Valley it found a broadly irregular surface with low hills and broad level spaces covered with a deep mantle of disintegrated rock. . . .

"The unconformity is well shown at Fort Mountain (Redoubt), where the basal Cambrian is formed of massive layers 4-10 feet thick, which usually rest directly on the Hector shale (pre-Cambrian). In places, however, slight hollows in the shale are filled with thin layers of a more or less ferruginous sandstone that was deposited by gentle currents prior to the deposition of the massive conglomerate layers. The lower 10-20 feet of this conglomerate contains rounded and angular fragments of the subjacent pre-Cambrian formations."

The present status of the problem may be briefly summarized. The Cambrian—pre-Cambrian contact in this part of Alberta is not conclusively established, as the strata of the two systems are generally conformable. The presence of intraformational conglomerates similar to those in the Cambrian rocks and containing white glassy quartz pebbles similar to those in the basal Fort Mountain sandstone suggests that the Hector may have been deposited in the older part of the Waucobian instead of in the Algonkian. However, if the Fort Mountain-Hector contact is the Cambrian—pre-Cambrian boundary, the amount of deformation and erosion at the end of the pre-Cambrian in this part of Canada was considerably less than in Montana (Deiss, 1935, p. 95-124).

MIDDLE-LOWER CAMBRIAN BOUNDARY

The first attempt to separate the Middle from the Lower Cambrian sediments in this part of Alberta and British Columbia was made by Walcott (1908a, p. 2-4), who assigned the Mount Whyte to the Lower and the Cathedral to the Middle Cambrian, thus indicating the contact between them as the boundary. In his summary-sections of Castle Mountain and Mount Bosworth, Walcott (1908a, p. 2) placed the Mount Whyte as the lowest Middle instead of the highest Lower Cambrian formation. Later the same year (Walcott, 1908c, p. 212) he published his Mount Bosworth section and again drew the Middle-Lower Cambrian line at the top of the Mount Whyte. In doing so he made the first specific statement concerning this boundary.

"The line between the Middle and Lower Cambrian is placed at this horizon on account of the presence in the Mount Stephen section of *Olenellus* in the limestone 116 feet below the massive arenaceous limestone belt represented by 1f in the Cathedral formation of the Mount Bosworth section."

In the Mount Bosworth section Walcott prematurely assigned *Albertella* to a horizon 255 feet below the top of the Mount Whyte. In doing so he said (Walcott, 1918c, p. 214):

"In the Lake Agnes section 5 miles southeast of Mount Bosworth, the fauna of about this horizon includes *Micromitra* (*Paterina*) *wapta* Walcott. . . . *Albertella*, s.p undt., *Bathyuriscus*. On the south slope of Mount Bosworth two drift blocks of siliceous shale, supposed to be from this horizon, were found, from which the following species were found:".

Albertella bosworthi Walcott, *A. helena* Walcott, and 13 other species were then listed from the two localities.

This assignment of the *Albertella* fauna to the Lower Cambrian was questioned by Burling (1914, p. 116-129) upon the paleontologic evidence that the fauna contained "forms either typical of the Middle Cambrian or confined to the *Albertella* fauna," and that none of the species had ever been found in association with *Olenellus*. Upon the thesis that a sudden and widespread introduction of a new fauna (in this case, the *Albertella* fauna) is of much more importance and significance than the ultimate extinction of an old fauna (*Olenellus*), Burling (1914, p. 129) proposed that the Middle-Lower Cambrian line be drawn at the base of the *Albertella* zone. As this zone was then undiscovered but believed to be in the Mount Whyte where Walcott placed it, Burling suggested that this formation was largely Middle instead of Lower Cambrian and that the boundary was somewhere in its lower part.

In 1916 Burling (1916a, p. 99; 1916b, p. 469-472) noted his discovery of the *Albertella* fauna several hundred feet above the top of Mount Whyte in the Cathedral formation on Castle Mountain and Mount Bosworth. Burling, however, still believed that "all but the lowest beds of the Mount Whyte formation are of Middle Cambrian age," but conceded that

> "the discovery that the *Albertella* fauna occurs in a shale member embedded 375 feet up in the overlying massive arenaceous limestones of the Cathedral formation robs me of the main arguments which I used in 1914 for the Middle Cambrian age of the Mount Whyte."
>
> (Burling, 1916b, p. 471.)

The following year Walcott (1917a, p. 1-4) defined his new Ptarmigan formation and placed the Middle-Lower Cambrian boundary at the Ptarmigan-Mount Whyte contact beneath the *Kochaspis chares* zone. Although Walcott erroneously assigned the Ross Lake shale with its *Albertella* fauna to the Ptarmigan formation, he correctly included such characteristic Middle Cambrian genera as *Kochaspis* in the Ptarmigan. Later the same year Walcott (1917c, p. 62-64) inconsistently placed *Kochaspis* and its associated genera in the part of the Mount Whyte and tentatively considered them to be Lower Cambrian. In so doing he said:

> "All of the stratigraphic sections in their upper portion have varying thicknesses of calcareo-arenaceous beds and bands of oolitic limestone. In the upper layers of limestone there is usually a well-marked fauna and at this horizon in three sections, at Mount Assiniboine, Ptarmigan Peak, and Mount Stephen, the fauna includes a representation of the genus *Crepicephalus*, which attains its greatest development in Middle and Upper Cambrian of America . . .
>
> "The fauna listed from the three localities give a general conception of the fauna associated with *Crepicephalus* in the upper portion of the formation. Its relations now appear to be more with the Mount Whyte than with the Ptarmigan fauna, but further field work may change this view."

In the faunal lists referred to in this quotation *Olenellus* and other characteristic Lower Cambrian genera do not appear. In his faunal lists

for the entire formation, however, Walcott (1917c, p. 64-65) indicated that *Olenellus canadensis* Walcott, *O. gilberti* Meek, and *O.* sp. undet. occur in the upper part of the Mount Whyte.

Burling apparently thought the assignment of *Olenellus* to the upper, middle, and lower Mount Whyte was correct but believed that this genus lived on into and became extinct early in the Middle Cambrian. Consequently, as late as 1922, he summarized the problem of the Middle-Lower Cambrian boundary by saying (Burling, 1922, p. 460-461):

"Dr. Walcott believes that the 'stratigraphic position of the Mount Whyte justifies its classification with the Lower Cambrian' and, I believe, that the stratigraphic relations near the base of the formation justify the transfer of its major and upper portion to the Middle Cambrian. Wherever the line is drawn we both seem to be agreed that the Lower-Middle Cambrian boundary can be determined by a study of the stratigraphy. Since the broader stratigraphy of a given bed is not so much a matter of opinion as of observation, the only inference that can be drawn is that there are breaks of more or less magnitude at both the top and bottom of the formation. In interpreting these Dr. Walcott refuses to admit the possibility of the recurrence in Middle Cambrian time of surviving remnants of the Mesonacidae; the writer sees in the sudden introduction of new forms evidence far more dependable than the ultimate extinction of the old."

Grabau recently discussed the Mount Whyte formation and correctly suggested that the Lower-Middle Cambrian boundary may be within the formation because *Kochaspis cecinna* (Walcott) is more diagnostic of the Middle than of the Lower Cambrian. In this connection Grabau (1936, p. 46) wrote:

"This is a fauna of rather remarkable composition, and one that cannot readily be considered as entirely typical of the Lower Cambrian. Indeed both the *Crepicephalus* and *Dorypyge* are generic types more characteristic of the Middle Cambrian and to that division the upper limestone of the Mount Whyte formation may perhaps be related. The lower part however, has an undoubted Lower Cambrian fauna and if a division is necessary, the dividing line must be placed somewhere within the formation, and this is probably within some of the sandstone members. It is true that the brachiopods of this fauna as well as the gastropods range widely through lower beds but the trilobites, with two exceptions do not occur below, these being *Crepicephalus cecinna*, which also occurs in one of the highest beds of the Mount Whyte of Ptarmigan Peak (1c) and the other in an upper bed (1b) of the Mount Bosworth section."

When the writer remeasured and emended the Highland Range section in Nevada (Deiss, 1938b, p. 1154-1160) he found that *Kochaspis augusta* (Walcott) and *K. liliana* (Walcott) are not associated with *Olenellus* but occur 338 feet above the highest horizon of *Olenellus* and at least 280 feet below the *Albertella* zone. From the facts that a significant lithologic change occurs beneath the *Kochaspis liliana* zone, that most of the species of *Kochaspis* elsewhere in the Cordilleran trough are of Middle Cambrian age, and that the genus is much more typically a Middle than a Lower Cambrian type, the writer drew the Middle-Lower Cambrian boundary in Nevada at the base of the lowest *Kochaspis* zone. When the Mount Whyte and Ptarmigan formations were remeas-

ured in 1938, extreme care was used to try to discover a break in the sequence that might represent the Middle-Lower Cambrian boundary.

The beds were seached for fossils, and each faunal zone discovered was collected separately and zoned as a separate fauna and fossil locality in the measured sections. It can now be definitely stated that *Olenellus* is abundant in the 24 to 26 feet of limestones above the base of the Mount Whyte, that doubtful and rare fragments of *Olenellus* occur 6 feet above the basal zone, but that not a trace of the genus could be found in any of the beds in the upper 243 feet of the formation. *Plagiura cercops* (Walcott), *Plagiurella cleadas* (Walcott), and other Lower Cambrian species occur 115 feet above *Olenellus* and 138 feet below the first *Kochaspis* zone in the base of the Ptarmigan limestone. Thus the faunal sequence in Alberta is surprisingly similar to that in Nevada.

In light of these facts, and upon the principle followed by Burling that the sudden appearance of a new fauna denotes a significant change of time and events, the Middle-Lower Cambrian boundary in Alberta and British Columbia is tentatively drawn below the first *Kochaspis* zone. The paleontologic argument for considering the first occurrence of *Kochaspis* in the Cordilleran region as an arbitrary marker of the beginning of the Middle Cambrian is greatly strengthened because *Kochaspis* is still well represented in the youngest Middle Cambrian fauna yet reported—the *Kochaspis upis* fauna of northwestern Montana (Deiss, 1939, p. 45).

On Castle Mountain and Ptarmigan Peak the only physical indication of the change from Lower to Middle Cambrian is the thin-bedded massive black-gray limestone which forms the base of the Ptarmigan formation. At Ross Lake the change is marked by the first of the tan-weathering medium-grained dolomites at the base of the Cathedral formation. An erosion surface or other evidence of a physical break between the two series could not be found, and none is believed to exist in the area. Consequently the evidence is not conclusive, and the horizon which has been indicated as the Middle-Lower Cambrian boundary must be thought of as the best working hypothesis now available.

UPPER-MIDDLE CAMBRIAN BOUNDARY

When Walcott first published the Mount Bosworth section (Walcott, 1908c, p. 204-217) he placed the Upper-Middle Cambrian boundary between the Eldon and Bosworth formations. Since then the boundary has never been changed or questioned. In the type section of his Eldon formation on Castle Mountain, Walcott (1928, p. 275) included the 580 feet of beds above the cliff-forming massive dolomite which are now placed in the Pika formation. From these beds Walcott reported only

"Annelid trails and two or three species of trilobites." However, an early Upper Cambrian fauna composed of numerous cranidia and pygidia of new genera occurs 93 feet above the base of the Pika. Further, the lithologic change from the underlying massive cliff-forming gray dolomite of the emended Eldon to the overlying thin-bedded limestones and shales of the Pika is greater than that from the Pika to the Arctomys (lower part of the original Bosworth formation of Walcott).

In light of this new evidence the Upper-Middle Cambrian boundary is placed 500 feet or more stratigraphically lower than it was previously drawn and lies between the Eldon and Pika formations on Castle Mountain (Fig. 2) and Ptarmigan Peak (Fig. 3), instead of between the Eldon and Bosworth (Walcott, 1908c, p. 208), or between the Eldon and Arctomys (Walcott, 1928, p. 275).

CORRELATION OF FORMATIONS

GENERAL STATEMENT

The Lower and Middle Cambrian formations in Alberta and British Columbia which were remeasured in 1938 are not exactly equivalent to any of the formations in the Cordilleran trough in the United States. However, the relative stratigraphic position and age of the formations in the different parts of the trough are now known from their contained fossils. The faunas in Canada can now be partly recognized in different parts of the trough to the south. The chronologic correlations given in the following pages are based upon these faunas.

LOCAL CORRELATIONS

The formations in the three sections remeasured in 1938 are correlated upon stratigraphic position, lithic characteristics, and available fossils. Their distribution within the area, and their peculiarities and fossils are given in connection with the definitions of the formations on pages 993 to 1010. The correlations of the formations within the area are graphically shown on Figure 5 and need not be discussed.

REGIONAL CORRELATIONS

General statement.—The regional correlation chart (Fig. 6) is based upon seven successive faunas, of which two are Lower and five are Upper Cambrian. These faunas are now recognized as the most valuable for correlation within the Cordilleran province because they have been found over wide distances, are readily recognizable and generally clearly defined, and are known to occupy the same relative stratigraphic position wherever found in this geosyncline. From oldest to youngest these faunas are:

1. The *Olenellus* fauna. The only zone in which this fauna was found

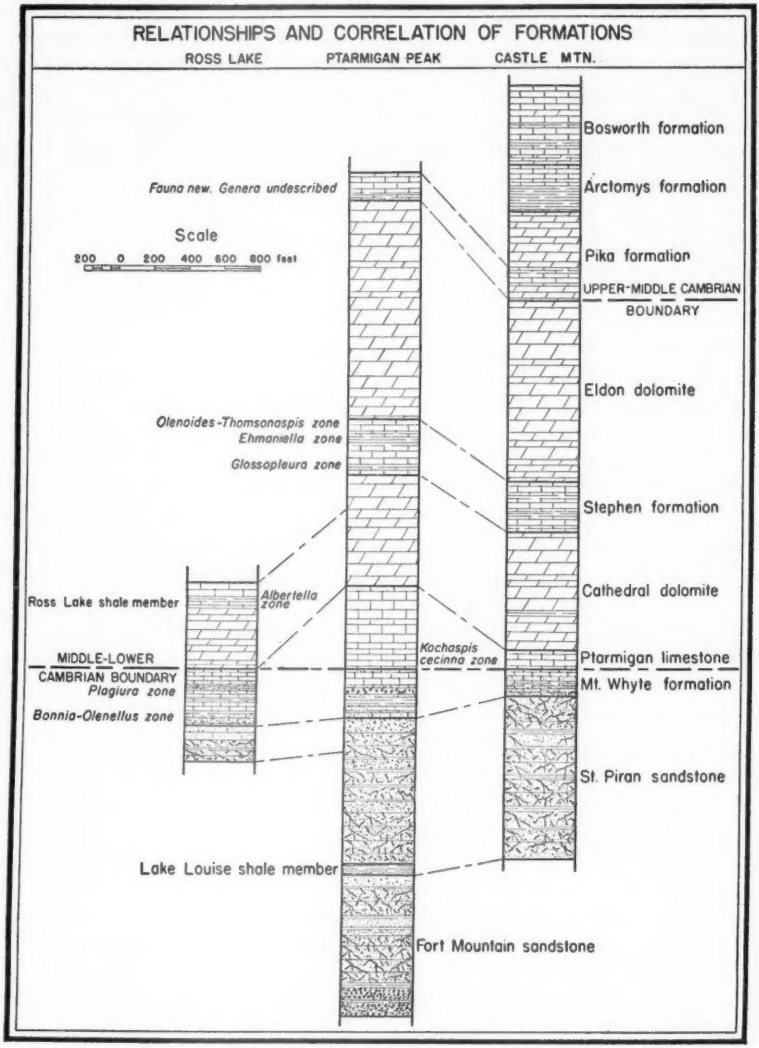


FIGURE 5.—*Relationships and correlation of formations in area*

is in the lower part of the Mount Whyte where it is most prolific from 24 to 26 feet above the base of the formation. *Bonnia* is represented in the fauna by numerous individuals.

2. The *Plagiura* fauna also occurs in the Mount Whyte, 115 feet above the *Olenellus* zone. This fauna is somewhat dwarfed and contains many individuals of *Plagiurella* and a few of several other genera.

3. The *Kochaspis* fauna is represented in Canada by *K. cecinna* (Walcott) and other species in the basal part of the Ptarmigan formation, 138 feet above the *Plagiura* zone (Fig. 3). The fossils in this fauna are poorly preserved and rare.

4. The *Albertella* fauna occurs 300 to 500 feet or more above the *Kochaspis cecinna* zone and is one of the most widespread and prolific Middle Cambrian faunas in the Cordilleran province.

5. The *Glossopleura* fauna in the basal part of the Stephen formation is approximately 500 feet above the *Albertella* zone on Castle Mountain and is another widespread and sharply delimited biota. The *Anoria-Clavaspidella* fauna of Montana (Deiss, 1938b, fig. 7) is not recognizable in Alberta but may be partly represented in the *Glossopleura* fauna.

6. The *Elrathina-Bathyuriscus-Agnostus* fauna of the Pentagon shale in Montana is represented in the *Ogygopsis* shale member of the Stephen formation on Mount Stephen but is absent from the other sections in Canada. As this fauna is older than the *Kochaspis upis* fauna which contains *Thomsonaspis* in Montana, and as *Thomsonaspis* occurs in the highest beds of the Stephen on Castle Mountain, the *Ogygopsis* shale fauna may be older than the upper part of the Stephen in Alberta.

7. The *Olenoides-Thomsonaspis* fauna is the youngest Middle Cambrian fauna in the Castle Mountain area and occurs in the uppermost Stephen, approximately 200 feet above the *Glossopleura* zone.

Fort Mountain and St. Piran sandstones.—The basal clastic sediments of the Lower Cambrian in southwest Alberta may be temporally equivalent to the Prospect Mountain quartzite and possibly to the lower part of the Pioche shale in the Highland Range of Nevada (Deiss, 1938b, p. 1157-1160). This correlation is suggested by the presence of *Olenellus gilberti* (Meek) 260 feet above the base of the Pioche in Nevada and in the basal beds of the Mount Whyte in Alberta. Similar forms of *Olenellus* occur in the Prospect Mountain quartzite in the type locality.[6] This correlation is of value only as a generalization and rests upon the inconclusive facts that in both areas the formations are as old or older than the last *Olenellus* zone and form the base of the Lower Cambrian in the two areas.

Mount Whyte formation.—*Olenellus gilberti* Meek occurs in the lower middle part of the Pioche shale in Nevada, in the upper part of the

[6] Dr. H. E. Wheeler, of the Nevada Bureau of Mines, recently sent to the writer fragmentary specimens of *Olenellus* cf. *gilberti* (Meek) from the upper part of the Prospect Mountain quartzite on Prospect Mountain in the Eureka District of Nevada (Hague, 1883, p. 254).

"Prospect Mountain quartzite" in Utah, and in the lower part of the Mount Whyte formation in Alberta and British Columbia. The *Plagiura* fauna occurs more than 100 feet above the *Olenellus* zone in the Mount

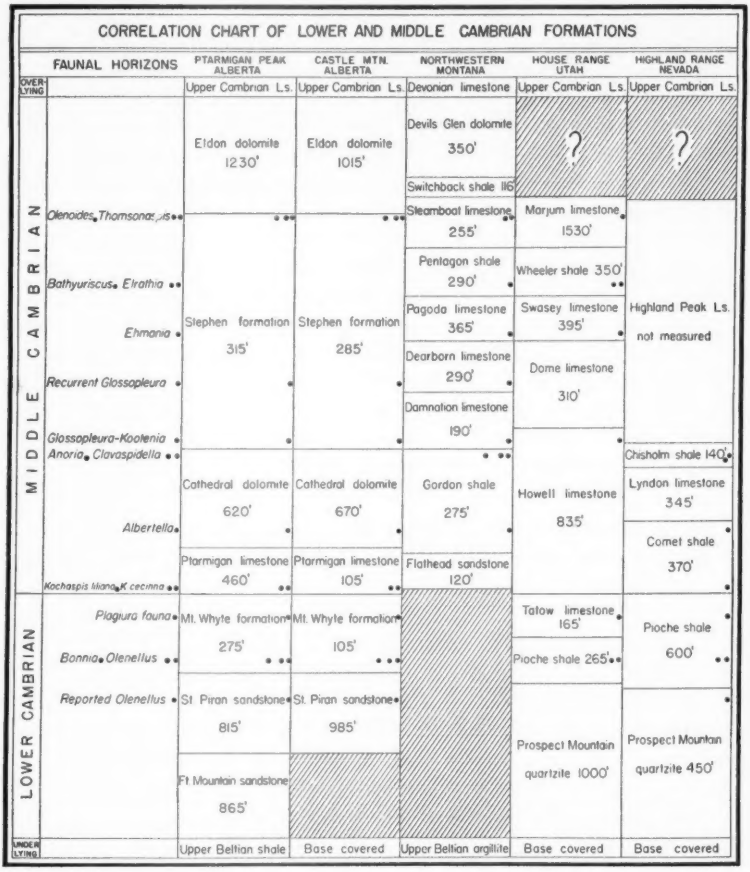


FIGURE 6.—*Regional correlation chart*

Whyte. Mason (1938, p. 290) reported the *Plagiura* fauna from the Highland Range, Nevada, but he erroneously associated it with the much younger *Kochaspis liliana* fauna. Another Lower Cambrian genus, *Poulsenia*, occurs 200 feet above the *Olenellus* zone in the House Range, Utah, in the Tatow limestone. Consequently, the Mount Whyte is tenta-

tively correlated with most of the Pioche shale of Nevada and with the Tatow limestone and the upper part of the Pioche shale in Utah.

Ptarmigan limestone.—The presence of such species as *Kochaspis cecinna* (Walcott) and *K. gogensis* (Walcott) in the basal part of the Ptarmigan limestone and of *K. liliana* (Walcott) and *K. augusta* (Walcott) in the basal beds of the Comet shale of Nevada is considered sufficient evidence for the correlation of the Ptarmigan with the lower part of the Comet. *Kochaspis* first appeared in the southern part of the Cordilleran geosyncline early in the Middle Cambrian in such species as *K. augusta* and *K. liliana*. At approximately the same time the genus first appeared in the Canadian part of the geosyncline in the lower part of the Ptarmigan limestone where it is represented by such species as *K. cecinna* and *K. gogensis*. Because this *Kochaspis* fauna is apparently absent in Montana, where the oldest known Cambrian fossils are in the lower part of the Gordon shale (Deiss, 1939, p. 37) and belong to the much younger *Albertella* fauna, the Ptarmigan limestone of Alberta may be of the same age as part or all of the Flathead sandstone in Montana. The Ptarmigan may also be equivalent to the lower part of the Howell limestone in the House Range of Utah, but evidence from fossils is not available. The only fossils in the Howell belong to the *Glossopleura* fauna and occur in the upper part of the formation (Deiss, 1938b, p. 1135-1138).

Cathedral dolomite.—The Ross Lake shale member in the middle part of the Cathedral contains the *Albertella* fauna, which in this area consists of 19 species. The entire fauna has never been found in any other region, but *Albertella helena* Walcott is reported from the lower part of the Gordon shale in northwestern Montana (Walcott, 1917b, p. 18), and new species of *Albertella* are reported from the upper part of the Comet shale in the Highland Range of Nevada (Deiss, 1938b, p. 1153). In all the known sections in which *Albertella* occurs the fauna has a short vertical range and lies below the *Glossopleura* zone. Because of these facts the Cathedral limestone may be correlated with the lower part of the Gordon shale of Montana and the upper part of the Comet shale of Nevada and may be equivalent in age to part of the Howell limestone in the House Range of Utah (Fig. 6).

Stephen formation.—The *Glossopleura* fauna occurs at the base, and the *Olenoides-Thomsonaspis* fauna at the top of the Stephen on Castle Mountain. Doubtfully identified species belonging to intermediate faunas in northwestern Montana are present in the medial beds of the Stephen. In northwestern Montana five distinct and widely separated faunas are present in 800 to 1200 feet of beds between and including the

Glossopleura and *Thomsonaspis* zones in the Stephen. The solution of this problem will require much additional field and laboratory work. However, several points are already well established.

The *Glossopleura* fauna is strikingly similar to the one in the base of the Damnation limestone in Montana (Deiss, 1939, p. 39) and to that in the upper part of the Howell limestone in the House Range, Utah. In these three areas the presence of such species as *Glossopleura fordensis* and others associated with similar species of *Kootenia* substantiate the correlation of the basal Stephen with the Damnation limestone of Montana and with the upper part of the Howell limestone of Utah. The presence of similar species of *Kootenia* in the basal Stephen and in the upper Wolsey shale of central Montana may indicate that these formations are partly equivalent.

Thomsonaspis occurs in northwestern Montana in the middle part of the Steamboat limestone where it is associated with *Kochaspis upis* (Walcott) and 13 other species and 5 genera, but *Olenoides* is not among them. In the upper part of the Stephen, *Thomsonaspis* is associated with *Olenoides* cf. *serratus* (Meek) and several species of *Solenopleurella* but with none of the genera or species known from this horizon in Montana. If the species of *Thomsonaspis* have a small vertical range, the upper part of the Stephen on Castle Mountain and Ptarmigan Peak is equivalent to the middle part of the Steamboat in northwestern Montana. *Olenoides* occurs in the upper part of the Marjum limestone in the House Range of Utah. Its presence there and in the upper Stephen may suggest the correlation of the upper parts of these formations.

In northwestern Montana between the *Glossopleura* and *Thomsonaspis* zones are the Dearborn, Pagoda, and Pentagon formations. The Stephen in Alberta apparently represents the time during which these formations were being deposited. The presence of closely similar species of *Agnostus, Bathyuriscus,* and *Elrathina* in the Pentagon shale of Montana and the *Ogygopsis* shale on Mount Stephen in British Columbia is considered conclusive evidence for the correlation of these shales. As neither the *Ogygopsis* shale nor its fauna is present in the Stephen formation on Castle Mountain, and as the exact faunal succession in the Stephen on Mount Stephen is unknown, the position of the *Ogygopsis* shale in the Castle Mountain area is undetermined.

In summary, the Stephen is equivalent to the upper part of the Howell limestone, to the combined Dome, Swasey, Wheeler and Marjum formations of Utah, and to the Damnation, Dearborn, Pagoda, Pentagon, and lower part of the Steamboat formations in northwestern Montana. Except for the conclusive evidence given by the *Glossopleura* fauna, the other correlations are tentative and are only partly supported by fossils.

Eldon dolomite.—Fossils are unknown from the emended Eldon formation, and the high magnesian content of the rocks precludes the probability that any identifiable ones will be found. Because the Eldon overlies the *Thomsonaspis* zone in Alberta, and the Switchback and Devils Glen formations are likewise above that zone in Montana, and because these formations are the youngest Middle Cambrian ones in both areas, the Eldon dolomite is tentatively correlated with the Switchback shale and the overlying Devils Glen dolomite in northwestern Montana.

The discovery of *Olenoides* associated with *Thomsonaspis* in the upper part of the Stephen and the tentative correlation of this zone with the *Kochaspis upis* zone in the Steamboat of northwestern Montana seem to imply that the Marjum may be equivalent to the middle Steamboat and therefore slightly younger than was known in 1938 when the writer correlated the Marjum with the Pentagon shale (Deiss, 1938b, p. 1163-1164, fig. 7). If the upper beds of the Marjum are of the same age as the upper Stephen and the middle Steamboat, this fact strongly supports the conclusion that the

> "Switchback and Devils Glen (and Eldon) were deposited in a northern sea after the water had withdrawn from the southern part of the Cordilleran trough in Nevada and western Utah."

(Deiss, 1938b, p. 1164.)

CONCLUSIONS

The results obtained from the study of the three sections and their contained fossils may now be summarized.

The original Mount Whyte formation contained both Middle and Lower Cambrian fossils. The upper 58 feet of the formation on Ptarmigan Peak contained *Kochaspis* cf. *cecinna* (Walcott), *K.* cf. *gogensis* (Walcott), and other Middle Cambrian species. The beds in this interval are lithologically identical to the overlying limestone and must be placed in the Ptarmigan formation, as they were originally (Walcott, 1917a, p. 1-4). The Middle Lower Cambrian boundary is then left at the base of the Ptarmigan formation below the *Kochaspis cecinna* zone where Walcott first placed it.

The youngest beds in the area which contain *Olenellus* are the basal 26 feet of the Mount Whyte, in which *Olenellus* is associated with *Antagmus?, Bonnia, Onchocephalus,* and *Syspacephalus.*

The Mount Whyte formation thins southward from 315 feet at Ross Lake and 275 feet at Ptarmigan Peak to 145 feet at Castle Mountain where it contains only a sparse fauna.

The Ptarmigan limestone is well developed only at its type locality on Ptarmigan Peak. There the formation consists of a black-gray lime-

stone which contains *Kochaspis* cf. *cecinna* (Walcott) and *K.* cf. *gogensis* (Walcott) in the basal beds and a new genus similar to *Kootenia* in the upper beds. The formation thins southward to 105 feet at Castle Mountain and is not recognizable either lithologically or faunally at Ross Lake.

The *Albertella* fauna, represented by seven species of *Albertella,* three of *"Amecephalus,"* two of *Vanuxemella,* and one each of *Kochina, Poliella, Ptarmingia, Hyolithes, Acrothele, Obolus,* and *Wimanella,* occurs only at one horizon in the Cathedral formation. The fauna is restricted to a shale 6 to 16 inches thick on Castle Mountain, 8 feet thick at Ross Lake, and 7 feet thick on Mount Bosworth. This shale is 200, 430, and 375 feet respectively above the base of the Cathedral in the three sections.

With the possible exception of *Albertella helena* Walcott, none of the species of trilobites in the upper part of the Gordon shale of Montana are associated with *Albertella* in Alberta or British Columbia. *Kochaspis, Olenoides,* and *Zacanthoides* also are not represented in the *Albertella* fauna in these sections in Canada and probably are not associated with *Albertella* in other areas.

The Stephen formation on Castle Mountain and Ptarmigan Peak contains genera restricted to thin successive zones which are equivalent to widely separated faunas in the Damnation, Dearborn, Pagoda, Pentagon, and Steamboat formations in northwestern Montana. The *Ogygopsis* shale fauna of Mount Stephen and the Burgess shale fauna of Mount Field are not present on Castle Mountain or Ptarmigan Peak. That strata containing these faunas may fill an unknown hiatus between the Stephen and Eldon in these two eastern sections is doubtful because this fauna is much older than the *Thomsonaspis* fauna which occurs at the top of the Stephen on Castle Mountain.

The Cathedral formation in the original Ptarmigan Peak section actually consists of 620 feet of Cathedral dolomite, 315 feet of Stephen limestone and shale, 1230 feet of Eldon dolomite, and 150 feet of the Pika (new) formation.

The Lower Cambrian strata in southwestern Alberta and southeastern British Columbia are divisible into three mappable formations on a scale of 1 mile to the inch. From the base upward these are: (1) Fort Mountain sandstone, (2) St. Piran sandstone, and (3) Mount Whyte formation. The Lake Louise shale is considered a member of the Fort Mountain sandstone.

In Alberta the Middle Cambrian strata are divisible into four mappable units: Ptarmigan limestone, Cathedral dolomite, Stephen formation (limestone and shale), and Eldon dolomite. The Ptarmigan limestone is not present in British Columbia in the vicinity of Ross Lake and Mount

Bosworth where it is either replaced by the Cathedral dolomite or fills an undiscovered hiatus between the Mount Whyte and overlying Cathedral.

The strata between the Eldon dolomite and Arctomys red shales contain an Upper Cambrian fauna, are readily mappable on a scale of 1 inch to the mile, and are named the Pika formation.

The Lower and Middle Cambrian sediments in the area are chronologically correlated with those in the Lewis and Clark Range of Montana, the House Range of Utah, and the Highland Range of Nevada. The correlations suggested are tentative and are based upon seven successive trilobite faunas.

WORKS TO WHICH REFERENCE IS MADE

Allan, J. A. (1911) *Ice River district, British Columbia,* Geol. Survey Canada, Summ. Rept. 1910, p. 135-144.

——— (1912) *Geology of Field map area, Yoho Park, B. C.,* Geol. Survey Canada, Summ. Rept. 1911, p. 175-187.

——— (1913) *Rocky Mountains (Bankhead to Golden),* Geol. Survey Canada, Guide Book no. 8, pt. 2, Transcontinental excursion C1, p. 167-201, map and structure section.

——— (1914a) *Geology of Field map area, B. C. and Alberta,* Geol. Survey Canada, Mem. 55, geol. ser., no. 46, 312 pages, map, 31 plates, 5 figures.

——— (1914b) *Rocky Mountain section between Banff, Alta., and Golden, B. C., along the Canadian Pacific Railway,* Geol. Survey Canada, Summ. Rept. 1912, p. 165-176.

——— (1917) *Geology of the Canadian Rocky Mountains,* Canad. Alpine Jour., vol. 8, p. 108-117.

Ashley, G. H., *et al.* (1933) *Classification and nomenclature of rock units,* Geol. Soc. Am., Bull., vol. 44, p. 423-459.

Burling, L. D. (1914) *Early Cambrian stratigraphy in the North American Cordillera, with discussion of Albertella and related faunas,* Geol. Survey Canada, Mus. Bull., no. 2, geol. ser., no. 17, p. 93-129.

——— (1916a) *Notes on the stratigraphy of the Rocky Mountains, Alberta and British Columbia,* Geol. Survey Canada, Summ. Rept. 1915, p. 97-100.

——— (1916b) *The Albertella fauna located in the Middle Cambrian of British Columbia and Alberta,* Am. Jour. Sci., 4th ser., vol. 42, p. 469-472, figs. 1, 2.

——— (1916c) *Stratigraphy of the Canadian Cordillera* (Abstract), Geol. Soc. Am., Bull., vol. 27, p. 158.

——— (1922a) *A Cambro-Ordovician section in the Beaverfoot Range, near Golden, British Columbia,* Geol. Mag., vol. 59, p. 452-461.

——— (1922b) *The relations between the Purcell Range and the Rocky Mountains in British Columbia, Canada,* Am. Jour. Sci., 5th ser., vol. 3, p. 254-256.

Camsell, Charles (1914) *Guide to the geology of the Canadian National Parks on the Canadian Pacific Railway between Calgary and Revelstoke,* Canada Dept. Interior, Nat. Parks Branch, Ottawa, p. 20-33, geol. map C1.

Daly, R. A. (1913) *Introduction to the geology of the Cordillera,* Geol. Survey Canada, Guide book no. 8, pt. 2, Transcontinental excursion C1, p. 111-166.

Dawson, G. M. (1891) *Note on the geological structure of the Selkirk Range,* Geol. Soc. Am., Bull., vol. 2, p. 165-176.

Dawson, G. M. (1901) *Geological record of the Rocky Mountain region in Canada,* Geol. Soc. Am., Bull., vol. 12, p. 57-92.

Deiss, Charles (1935) *Cambrian-Algonkian unconformity in western Montana,* Geol. Soc. Am., Bull., vol. 46, p. 95-124, pls. 7, 8.

——— (1936) *Revision of type Cambrian formations and sections of Montana and Yellowstone National Park,* Geol. Soc. Am., Bull., vol. 47, p. 1257-1342, 2 plates, 10 figures.

——— (1938a) *Middle Cambrian trilobite faunas in the southern part of the Cordilleran Trough* (Abstract), Geol. Soc. Am., Pr. 1937, p. 274.

——— (1938b) *Cambrian formations and sections in part of Cordilleran Trough,* Geol. Soc. Am., Bull., vol. 49, p. 1067-1168, 2 plates, 7 figures.

——— (1939) *Cambrian formations and sections in northwestern Montana,* Geol. Soc. Am., Spec. Paper no. 18, 135 pages, 18 plates, 7 figures.

Evans, C. S. (1933) *Brisco-Dogtooth map area, British Columbia,* Geol. Survey Canada, Summ. Rept. 1932, pt. A2, p. 106-176.

Grabau, A. W. (1936) *Palaeozoic formations in the light of the Pulsation theory,* Univ. Press, Nat. Univ. Peking, vol. 1, 2d ed., p. 39-64, 219-232, 236-251.

——— (1937) *Palaeozoic formations in the light of the Pulsation theory,* Univ. Press, Nat. Univ. Peking, vol. 3, pt. 2, p. 211-281.

Hague, Arnold (1883) *Abstract of the report on the geology of the Eureka District,* U. S. Geol. Survey, 3d Ann. Rept., p. 254.

Kobayashi, Teiichi (1936) *Cambrian and Lower Ordovician trilobites from northwestern Canada,* Jour. Paleont., vol. 10, p. 157-167.

Mason, J. F. (1938) *Cambrian faunal succession in Nevada and California,* Jour. Paleont., vol. 12, no. 3, p. 287-294.

McConnell, R. G. (1887) *On the geological structure of a portion of the Rocky Mountains,* Geol. Survey Canada, Ann. Rept. 1886, n. s., vol. 2, pt. D, 41 pages.

——— (1889) *Note on the geology of Mt. Stephen, British Columbia,* Am. Geol., vol. 3, p. 22-25.

——— (1896) *Report on an exploration of the Finlay and Omenica rivers,* Geol. Survey Canada, Ann. Rept., vol. 7, pt. C, p. 24c, 25c.

Raymond, P. E. (1930a) *The Paleozoic formations in Jasper Park, Alberta,* Am. Jour. Sci., 5th ser., vol. 20, p. 289-300.

——— (1930b) *The base of the Ordovician in the Canadian Rockies,* Am. Jour. Sci., 5th ser., vol. 20, p. 301-307.

——— (1931) *A structure section across the Canadian Rockies,* Jour. Geol., vol. 39, no. 2, p. 97-116, figs. 1-3.

Resser, C. E. (1933) *Preliminary generalized Cambrian time scale,* Geol. Soc. Am., Bull., vol. 44, p. 735-756.

———, and Bridge, Josiah (1932) *Stratigraphic nomenclature in the United States,* 16th Intern. Geol. Congr., U. S. 1933, Guidebook 29, pl. 1.

Rominger, Carl (1887) *Description of primordial fossils from Mount Stephens, N. W. Territory of Canada,* Philadelphia Acad. Nat. Sci., Pr. 1887, p. 12-19, pl. 1.

Schofield, S. J. (1914) *The pre-Cambrian (Beltian) rocks of southeastern British Columbia and their correlation,* Geol. Survey Canada, Mus. Bull., no. 2, geol. ser., no. 16, p. 79-91.

——— (1920) *The origin of the Rocky Mountain Trench, British Columbia,* Royal Soc. Canada, Pr. Tr., 3d ser., vol. 14, sec. 4, p. 61-97, 5 figures.

——— (1922) *Relationship of the pre-Cambrian (Beltian) terrain to the Lower Cambrian strata of southeastern British Columbia,* Geol. Survey Canada, Bull., no. 35, geol. ser., no. 42, p. 3-15.

Walcott, C. D. (1888) *Cambrian fossils from Mount Stephens, Northwest Territory of Canada,* Am. Jour. Sci., 3d ser., vol. 36, p. 163-166.

——— (1890) *The fauna of the Lower Cambrian or Olenellus zone,* U. S. Geol. Survey, 10th Ann. Rept., pt. 1, p. 515-629, pls. 63-98, figs. 44-69.

——— (1908a) *Nomenclature of some Cambrian Cordilleran formations,* Smithson. Misc. Coll., vol. 53, p. 1-12.

——— (1908b) *Cambrian trilobites,* Smithson. Misc. Coll., vol. 53, p. 13-52,

——— (1908c) *Cambrian sections of the Cordilleran area,* Smithson. Misc. Coll., vol. 53, p. 167-230, pls. 13-22.

——— (1908d) *Mount Stephen rocks and fossils,* Canad. Alpine Jour., vol. 1, no. 2, p. 232-248, 8 plates.

——— (1910a) *Pre-Cambrian rocks of the Bow River Valley, Alberta, Canada,* Smithson. Misc. Coll., vol. 53, p. 423-431, pls. 45-47.

——— (1910b) *Olenellus and other genera of the Mesonacidae,* Smithson. Misc. Coll., vol. 53, p. 231-422, pls. 23-44.

——— (1912a) *Cambro-Ordovician boundary in British Columbia with description of fossils,* Smithson. Misc. Coll., vol. 57, p. 229-237.

——— (1912b) *Cambrian Brachiopoda,* U. S. Geol. Survey, Mon. 51, 872 pages, 104 plates.

——— (1912c) *Cambrian of the Kicking Horse Valley, B. C.,* Geol. Survey Canada, Summ. Rept. 1911, p. 188-191.

——— (1913) *Cambrian formations of the Robson Peak District, British Columbia and Alberta, Canada,* Smithson. Misc. Coll., vol. 57, p. 327-343, pls. 55-59, figs. 11, 12.

——— (1915) *Cambrian problems in the Cordilleran region,* Problems of American Geology, Yale Univ. Press, New Haven, Conn., p. 162-233.

——— (1916) *Cambrian trilobites,* Smithson. Misc. Coll., vol. 64, p. 303-456, pls. 45-67.

——— (1917a) *Nomenclature of some Cambrian Cordilleran formations,* Smithson. Misc. Coll., vol. 67, p. 1-8.

——— (1917b) *The Albertella fauna in British Columbia and Montana,* Smithson. Misc. Coll., vol. 67, p. 9-59, 7 plates.

——— (1917c) *Fauna of the Mount Whyte formation,* Smithson. Misc. Coll., vol. 67, p. 61-114, pls. 8-13, figs. 1-3.

——— (1923) *Nomenclature of some post-Cambrian and Cambrian Cordilleran formations,* Smithson. Misc. Coll., vol. 67, p. 457-476, fig. 24.

——— (1927) *Pre-Devonian sedimentation in the southern Canadian Rocky Mountains,* Smithson. Misc. Coll., vol. 75, p. 147-173.

——— (1928) *Pre-Devonian Paleozoic formations of the Cordilleran provinces of Canada,* Smithson. Misc. Coll., vol. 75, p. 185-367, pls. 26-108, figs. 24-35. (Written by C. E. Resser after Walcott's death, from Walcott's notes and publications.)

Walker, J. F. (1926) *Geology and mineral deposits of Windermere map-area, British Columbia,* Geol. Survey Canada, Mem. 148, geol. ser., no. 129, p. 21-24.

Warren, P. S. (1929) *Sedimentary record in the Rocky Mountains at about the 51st parallel,* Canad. Field-Nat., vol. 43, no. 2, p. 23-25.

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